Master Thesis in Geosciences

Reservoir characterization of a fluvial sandstone

Depositional environment and heterogeneities in modeling of the Colton Formation, Utah

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FACULTY OF MATHEMATICS AND NATURAL SCIENCES

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Abstract

The aim of the present study has been to gather and handle data from a sandstone outcrop as an analogue for a hypothetical fluvial sandstone reservoir. Reservoirs generally have a limited set of data and number of data points. By using records of alluvial style, architecture and heterogeneities from outcrops, supplementary data are obtained for reservoir modeling.

The study object was the Paleocene/Eocene Colton Formation in the Roan Cliffs in Utah. The Colton Formation was a clastic wedge interchanging with and pinching out into the lacustrine Green River Formation at the southern margin of the Uinta Basin, reflecting progradation and later retrogradation of a coarse-clastic alluvial system from the east or southeast. Sandstone percentage and architecture are linked to base level fluctuations, depositional environment on the alluvial plain and rate of accommodation versus rate of sedimentation (A/S).

During depositional time of the Colton Formation, rate of accommodation was at first high and later slowly rising. Units of lacustrine facies or sediments deposited in a distal alluvial plain setting formed during events of rise in the lake level. Overlying medial alluvial plain sediments were initiated by base-level falls leading to erosion and subsequent slow base level rise and deposition of amalgamated multistorey and multilateral channel-belt sandstone bodies with high connectedness ratio. Increasing A/S ratio further upwards led to increasing preservation of floodplain fines and lower connectedness of channel sandstone bodies, reflecting a change from medial towards distal alluvial plain environment.

Data from the Colton Formation outcrop and from a well in the area are applied in modeling of sandstone architecture performed on Petrel software tool. On a macro scale, the modeling shows a high sandstone body connectedness above the datum. However, the mesoheterogeneity of the channel sandstone bodies is complex due to mud chip lags and mudstone drapes.

Acknowledgement

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Ane Rasmussen

List of content

1 Introduction

Fluvial reservoirs are characterized by rapid lateral and vertical facies changes. Well data gives limited information on the three dimensional extent of sandstone bodies, and well to well correlation is complicated. In analogue studies, architectural style, sandstone body geometries and heterogeneities are studied in outcrops. The resulting quantitative data and conceptual understanding can enhance the understanding of reservoir rocks.

The Paleocene/Eocene Colton Formation was a clastic fluvial wedge interchanging with and pinching out into the lacustrine Green River Formation at the southern margin of the Uinta Basin in Utah. The outcrops of the Colton Formation in the Roan Cliffs in Utah are well suited for studying large-scale development of fluvial architecture and heterogeneities.

The scope of this study was to gather field data on facies, facies associations, architectural style and heterogeneities from the Colton Formation. Data was used to analyze depositional environment and factors controlling deposition and alluvial architecture. A three dimensional model of the Colton Formation was generated based on data from the outcrop, the SAFARI project and a well drilled in the Minnie Maud canyon for coal-prospecting purposes.

2 Previous studies

Outcrop data collections from assumed reservoir analogues are performed for several purposes, here summarized by Dreyer et al. (1993):

- Reduce uncertainty in well-to-well correlation at a stage where well data are limited
- Create terms of reference for conceptual geological modeling
- Generate basic data for geomathematical modeling for reservoirs

Field studies for obtaining data for analogue studies are preferentially performed in areas where the advantage of lateral sandstone body correlation and connectedness in outcrop can be fully explored. Analogue studies have been performed in the Colton Formation by among others Taylor and Ritts (2004), the SAFARI project (SAFARI, 1995) and Morris et al. (1991, 1992). These studies are given a short description below.

2.1 Background studies

Taylor and Ritts reservoir analogue studies

The study of Taylor and Ritts (2004) utilized logging and photomosaic of outcrops of the Colton and Green River Formations to address the facies geometry and lateral heterogeneity. They also sampled sandstone specimens for petrographic analysis to further investigate the small-scale heterogeneity within sandbodies. Emphasis was put on the use of photomosaic for completion of mapping and facies analysis. The aim of their study was to investigate the complexity of sandstones in a lacustrine setting.

SAFARI project

Studies focusing on fluvial sedimentary rocks have been performed extensively by the SAFARI (Sedimentary Architecture of Field Analogues for Reservoir Information) project, a joint effort between Norsk Hydro, Norwegian Petroleum Directorate, Saga Petroleum and Statoil in the period 1989-1994. Focus of the project was towards improved reservoir characterization and it was executed as a cooperative field work resulting in a data base consisting of 2D outcrop maps, or *panels*. Computerized panels were available as databases for the participants for further reservoir concept development, as for instance reservoir analogue modeling purposes.

Analogues for fluvial reservoirs assumed to be generally applicable were collected from several localities by the SAFARI project. One of these was studies of the Colton Formation in 1988 to 1995. The Colton Formation was initially assumed to be an analogue to the upper reservoir zones of the upper member of the Lunde Formation and the Raude Member of the Statfjord Formation, based on parameters like grain size, sand: gross, facies, heterogeneity and depositional environment. Outcome of the Safari Project in the Colton Formation was 17 panels and 20 logs. Panels display alluvial architecture and heterogeneity, whereas additional lithological information can be found in the logs. Other SAFARI analogues that have led to several published articles are from the Escanilla and Sobrarbre formations in the Spanish Pyrenees (Dreyer et al., 1993).

Morris et al. field studies

The field studies of the Colton Formation performed by Morris et al. (1991 and 1992) were not presented as analogue field studies, but the in last papers discussed both heterogeneities and reservoir properties.

Discrepancies in interpretation of depositional environment within the Colton Formation between the present study and previous studies are few. In the SAFARI project the extensive, thin tabular sandstone bodies in the lower part of the formation were inferred to be sheet floods (Nystuen pers. communication, April 2004), whereas Morris et al. (1991, 1992) suggested them to be distributary mouth bars. In this study it is argued that these sandstone sheets are beach deposits, of which the uppermost one chosen is chosen as datum. Still the facies of the tabular sandstone bodies within and below the datum level are by no means unambiguous, and will be further discussed in Chapter 6.

3 Methods and data

3.1 Field work, available data and handling

Description of facies, facies associations and panels in this paper are based on data gathered during three weeks of field work in April 2004 in the Roan Cliffs in Utah. Participants of the field work were Katrine Brinck and Ane Rasmussen supervised by Johan Petter Nystuen of UIO.

Overall description and interpretation of depositional environment and modeling work in addition incorporates unpublished data from the SAFARI project and logs from a well drilled in Minnie Maud Canyon in 1994 for coal prospecting in the Mesa Verde Group. Modeling based on the data is performed on Petrel software tools.

3.2 Field data

Six logs were obtained in the Roan Cliffs in a 900 meter of continuous outcrop (Fig. 3.1 and 3.2). 2D panels were drawn in the field. The data set comprises 6 sections ranging from 42-379 meters in stratigraphic thickness in the panels 17 (logged by the author) and 18 (logged by Katrine Brinck), in total 1730 meters of strata. Logs are presented in Appendix and panels in Fig. 3.3.

The availability and conditions of outcrops are similar for this study and the SAFARI project, as they all are localized along the Roan Cliffs in Whitmore Park. The Roan Cliff is facing south in front of a marked hill between the valley of the Whitmore Park and the Indian Canyon to the north. Sedimentary rock cover in the area has a general dip of 007/350. The outcrop quality is generally good, and the lateral control of the thick sandstone units in the formation along the Roan Cliff is excellent. Clay and silt dominated intervals are generally covered by loose debris. Outcrop quality is best in the central, steepest part of the Roan Cliff, as sections below are for parts covered by scree, and outcrops close to the top of the hill are deeply weathered and eroded (figures 3.1 and 3.2). Lateral field correlation by walking along units was made in the central parts of the outcrop.

Data from the SAFARI project consists of logs and 2D panels taken from cliffs along the uppermost part of the Nine Mile Canyon and the Roan Cliffs, extending about 5 kilometers eastwards from the cliff logged in this study. In addition, the well provided data from a location about 5 kilometers north of our study area. All locations and model outline are shown on Fig. 3.2.

Figure 3.1.: Outcrop of Colton Formation presented in this study. Length of outcrop is 900 meters.

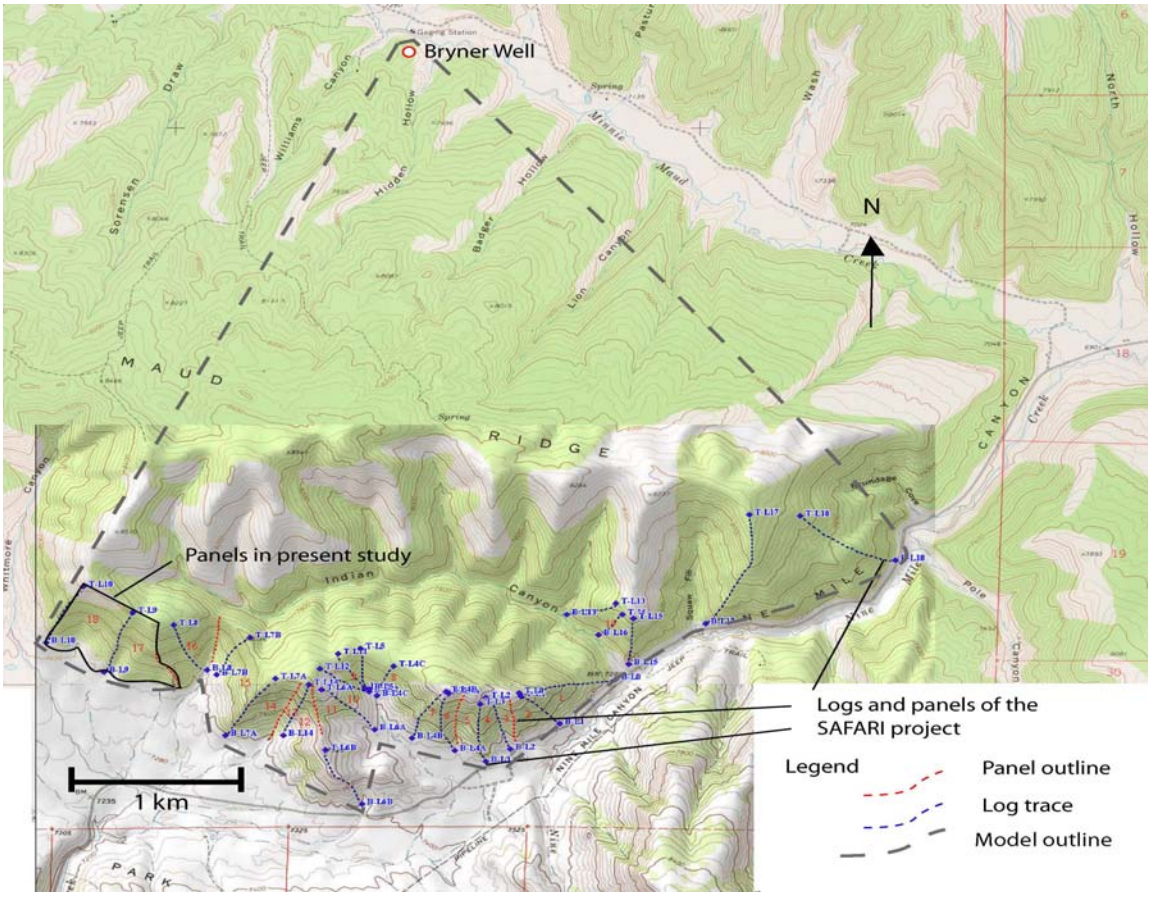


Figure 3.2: Map of geographical extent of data collection areas in Roan Cliffs.

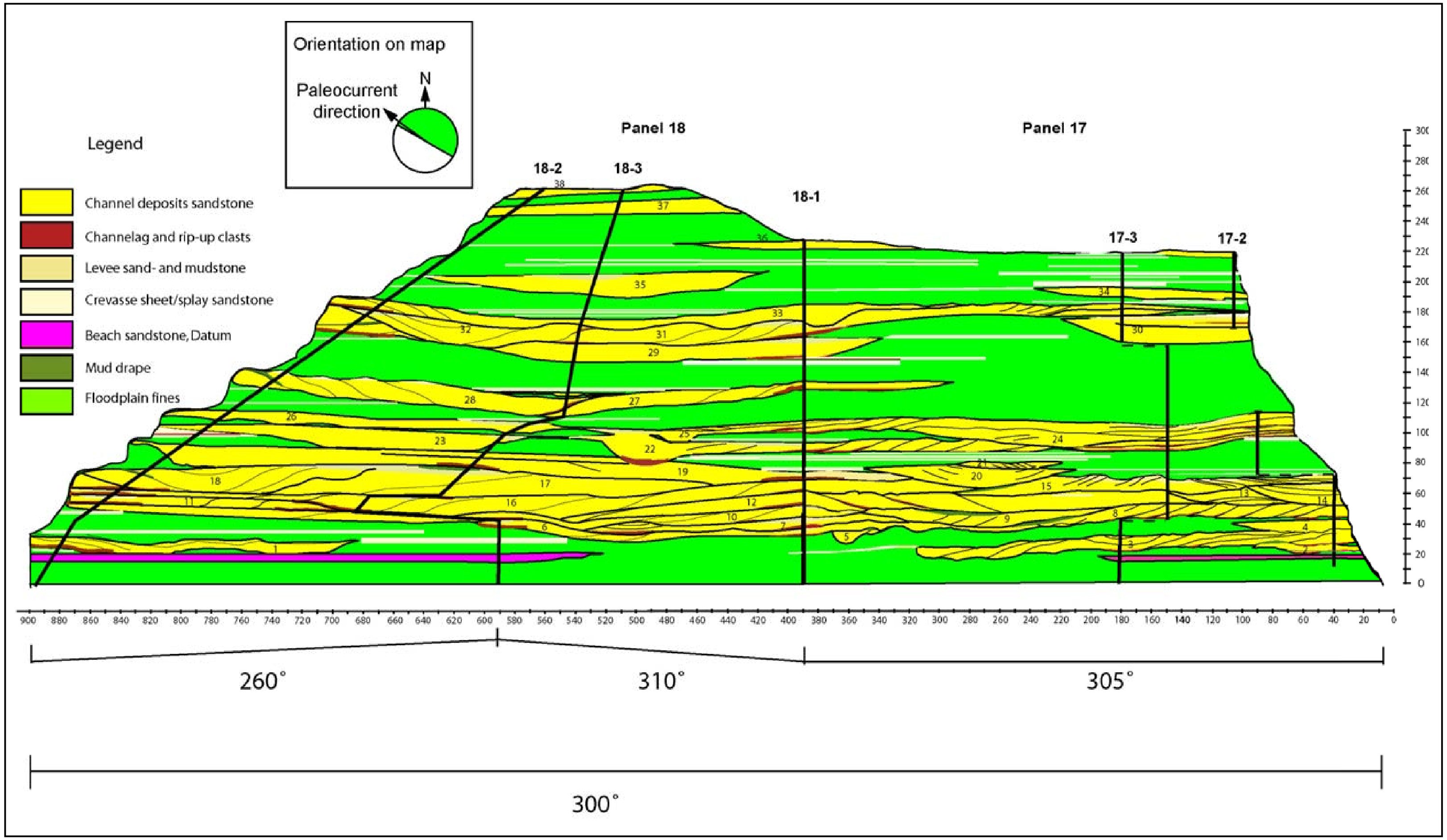


Figure 3.3: Panel 17 and 18. The panel is a 2D projection of the outcrop in the Roan Cliffs drawn. Panel starts at the datum. Actual logged outcrop height is over 370 meters. Log traces are marked on the panel, except for log 17-1 which is logged in a small hill in the easternmost part of panel 17. The top of log 17-2 is below the datum.

4 Geological setting

The Colton Formation is situated in the Paleocene/Eocene Uinta Basin on the Colorado Plateau in the north-eastern corner of Utah (Fig. 4.1). The Colorado Plateau is an uplifted area bounded in the west and south by the Basin and Range Province, in south-east by the Rio Grande Rift and in east and north by uplifted blocks and continental basins within the Rocky Mountains. It is a structurally stable block surrounded by tectonically active areas. The Colorado Plateau has a more or less continuous sedimentary record dated from the Precambrian and has today a large, immature erosional relief because of uplift that started 15 million years ago.

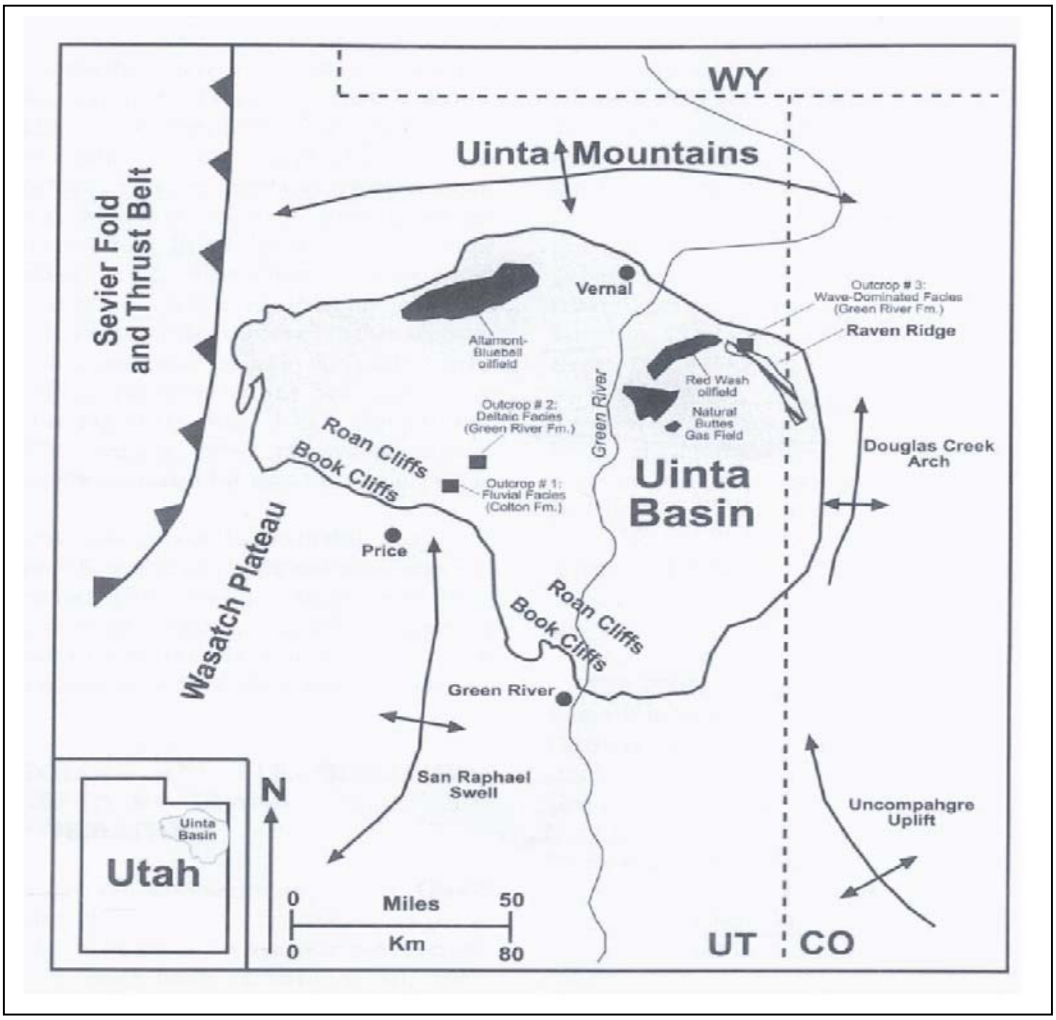


Figure 4.1: Paleogeographic map of the Uinta basin in Utah (Taylor & Ritts, 2004).

4.1 Geological development of the Colorado Plateau

During the geological history of the Colorado Plateau several structural features have been repeatedly reactivated. The formation of the Uinta Basin is linked to these reactivated structures. The brief summary of the geological development of the Colorado Plateau region given below is primarily sourced from Stokes (1988).

The present-day Uinta Mountains are located to south of and aligned with the Precambrian Archaean Craton of basements rocks, and were the area of an aulacogen in the Precambrian (1000-800 my) previous to the generation of the so called Wasatch Line. The Wasatch Line is a flexural-fault zone extending from north-east to south-west in Utah. This is an old zone of crustal weakness that has been repeatedly reactivated since the Precambrian (after 800 my). The Uncomphagre structural block is another feature that has been repeatedly reactivated by uplifts to a highland area and has been a source of clastic sediments to adjacent sedimentary basins. As a part of regional intercratonic uplift, extending from Utah to Oklahoma, it was first activated in the Mississippian – Permian.

In the Triassic to Early Cretaceous the Wasatch Line defined the boundary between continental and shallow marine platforms in the east and a deeper, marine environment in the west. The present day Colorado Plateau developed into an intracratonic basin, with deposition of aeolian, fluvial, lacustrine and some marine sediments. The marine transgression in the area came from the north. Further west, in present day Nevada and California, the orogenic belt of the Nevada Orogeny were formed due to collisions between the North American plate in the east and a Pacific oceanic plate in the west. During the late Cretaceous, the Western Interior Seaway developed as an epicontinental sea, extending from present day Gulf of Mexico to arctic Canada. The seaway was bounded to the west by the Sevier Orogenic Belt. Development of the Sevier Orogenic Belt was caused by thin-skinned tectonics east of the former Nevada Orogenic Belt as a result of a further development of the collision zone between the North American plate and the Pacific oceanic plate. The Western Interior Seaway formed a wide foreland basin along the eastern marginal zone of the Sevier Orogenic Belt. In the late Cretaceous the Western Interior basin was filled by sediments and gradually cut off from the ocean. The transition from Cretaceous to Tertiary was in central western North America characterized by a structural and sedimentological change from a wide marine environment to continental settings.

During late Cretaceous and Paleocene time the tectonic regime turned from thin-skinned tectonics in the Sevier Orogenic Belt to thick-skinned tectonics during the Laramide movements further to the east. These crustal movements started in the Campanian, while there still was ongoing thin-skinned tectonics in the west within the Sevier Orogenic Belt. A compressive structural regime in the Laramide Orogeny (also called the Rocky Mountain Orogeny) formed several small and large blocks or uplifted areas, elongated and mainly oriented north-north-west – south-south-east. The uplifted blocks are usually asymmetric with low-angled listric faults on the steep side, with the listric faults extending down to a depth of 15-20 kilometers. Several continental intermontane basins formed by subsidence of crustal blocks down to depths of up to several thousand meters. The basins were filled in by clastic material from the basin sides, while the central parts usually had lacustrine sedimentation; silt, mud, carbonates and organic matter.

The Uinta Basin is one of the Laramide continental basins. As the east-west oriented Uinta Mountains were uplifted as an elongated anticline, sedimentary basins developed in crustal depressions north and south of the anticlinal ridge. These basins are commonly known as the Green River Lakes (Stokes, 1988); in the north Green River (Lake Gosiute) and Washakie Basins and in the south Uinta Basin (Lake Flagstaff and Lake Uinta) and the Piceance Basin (Lake Piceance), see figure 4.2. In the area of the Uinta Basin, the San Rafael Swell, Monument Uplift and Circle Cliffs were also uplifted as large Laramide anticlines, and the Uncomphagre block was reactivated and uplifted. Erosion of the uplifted areas

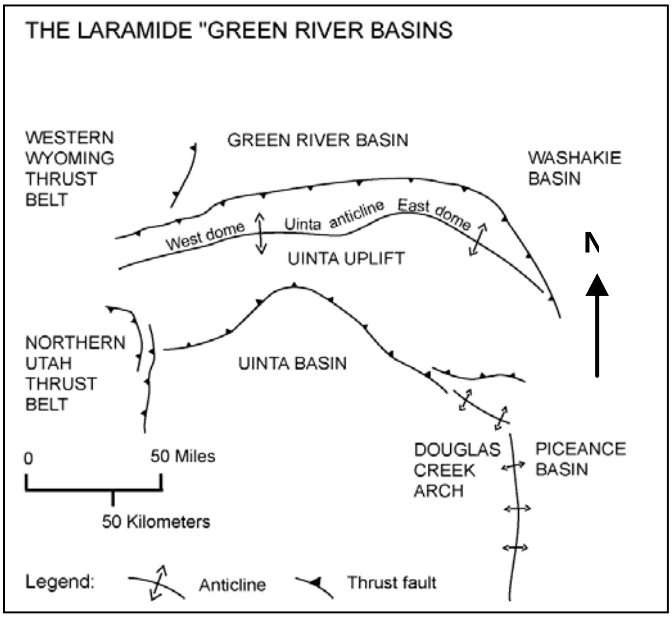


Figure 4.2: The Paleocene/Eocene Uinta Mountains and surrounding basins (Modified from Stokes, 1988)

provided ample sediments for the basin infill. The structurally controlled basins should ideally be oriented with the long axis north - south, as is seen in the precursor for the Uinta Basin; the Lake Flagstaff. Lake Flagstaff is a basin situated to the south-east of the later Uinta Basin. Stanley and Collinson (1979) proposed that the Lake Flagstaff was actually the final stage of infill of the foreland basin, and therefore was controlled by the vergence of the Sevier Orogeny instead of the Laramide Orogeny. The reactivation of former structural features, as the Uncomphagre block and, most likely, the Uinta Mountains, caused formation of basins with orientations controlled by the older structures. Fig. 4.3 shows the uplifts in the eastern Utah regarded to be results of the Laramide Orogeny, where the deviating orientation of the Uncomphagre and Uinta Uplift is seen.

4.2 The Uinta Basin

The sedimentary record of the Uinta Basin is dominated by sediments deposited in Lake Flagstaff and Lake Uinta. Lake Flagstaff was situated to the south-west of the later Lake Uinta (Fig. 4.3). It is debatable whether there actually was an overlap and open connection between these lakes; the lacustrine sedimentary rocks of Lake Flagstaff and Lake Uinta is usually distinct and can be separated in areas where Lake Uinta sediments (Green River Formation) is overlying Lake Flagstaff sediments (Flagstaff Formation). The Lake Flagstaff was after a while displaced westwards by the increased influx of clastic sediments from south and south-east and with time it also decreased in size (Smith, 1986). In the middle Paleocene, the

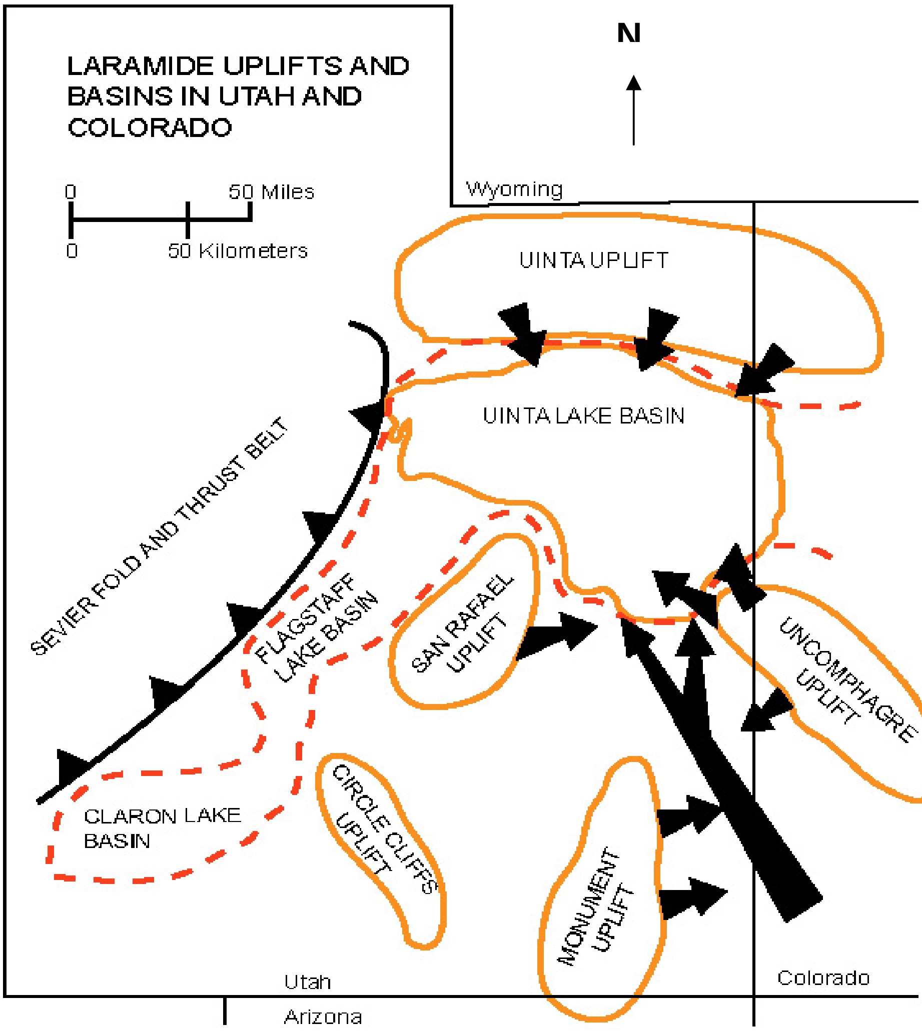


Figure 4.3: Paleocene/Eocene uplifts and basins of Utah (Modified from Stokes, 1988) showing provenance areas and depositional directions (arrows) for the Uinta Basin (Modified from Fouch et al, 1991). The approximate extent of Lake Flagstaff, Lake Uinta and the south-western Lake Claron shown as a dotted red line (Modified from Stokes, 1988).

lacustrine environment expanded again, and Lake Uinta was formed. The Piceance Basin was located east of Lake Uinta, being separated from the Lake Uinta by the Douglas Creek Arch, as seen in figure 4.2.

Lake Uinta had an area of about 20 000 km² (Picard and High, 1972). It was bounded in the west by the Sevier Thrust Belt and in the north by the Uinta Mountains. The southern and eastern basin margin slope and bottom dipped very gently towards north. The northern and western shores, bounded by high mountains, may have been fairly steep. Coarse clastic material, being deposited from the basin margins into the basin as fluvial deposits, deltas or alluvial sediments, built out as short wedges in the north and west and extended further out into the basin from the east.

The Lake Uinta was an asymmetric basin which was deepest in the north and northwest. Tectonic subsidence was taking place all around the Uinta Basin, but the northern and north-western subsidence was dominating, as can be concluded from the basin shape and infill. As well as tectonic subsidence on the southern boundary of the Uinta Mountains, there was movement on the Wasatch Line in the Paleocene causing a 20 km sinistral slip in the western Uinta Mountains (Picard et al. 1984).

The fossil record indicates a subtropical climate in the early and middle Eocene in the region of the Uinta Lake (Leopold and MacGinitie, 1972), with an average temperature of 6-18 ºC in the coldest month. According to their studies, there were no dry seasons, but a period in the earliest middle Eocene might have had seasonal drought. Later studies, as by Wilf et al. (1999) have suggested that there was a temperature increase from the late Paleocene to the early Eocene and later increased seasonality of rainfall from the early to the middle Eocene.

The Uinta Lake was probably not large enough to be subjected to tidal powers. Wind could rework the shoreline deposits, but the low gradient in the south-east would decrease the wave energy significantly before it reached the eastern shore. The climate and basin geometry makes it likely that the deltas building out into the lake from the east and south east were irregular, marshy and vegetated, without any significant reworking of the delta front sediments. Relatively low fluctuations in lake level can have had a prominent effect.

The Uinta Basin succession consists of continental and lacustrine sedimentary rocks. The lowermost formation is the Paleocene fluvial/lacustrine North Horn Formation, followed by the Eocene lacustrine Flagstaff Formation, the fluvial Colton Formation and the lacustrine Green River Formation, and, finally, the lacustrine/fluvial Uinta Formation (Fig. 4.4).

There are large lateral variations in the Uinta Basin succession regarding sedimentary facies. The boundaries between the formations are expressions of the dynamic interplay of factors such as tectonism, lake-level changes, palaeotopography, climate and time. A transect south to north across the Uinta Basin is shown in figure 4.5.

The North Horn Formation is a composite formation of the sedimentary rocks, mostly fluvial and lacustrine, overlying the Cretaceous Mesa Verde Group in the area. Some places the transition is gradual, but mostly there is a hiatus at the lower boundary of the North Horn Formation. The hiatus is an angular unconformity caused by deposition in the newly formed basin due to the Sevier tectonic movements.

The Flagstaff Formation is composed of lacustrine deposits, as limestone and shale. Sedimentary records reflect periods of shallow fresh water lake conditions alternating with playa conditions (Stanley & Collinson, 1979). In the southern and eastern parts of the basin clastic input was from source areas in the north and west.

The Colton Formation is clearly overlying the Flagstaff Formation in many parts of its area of extent (Fig. 4.6), but Colton Formation pinches out to the west and north, and Flagstaff pinches out to the east. In the western and northern areas around Soldier Summit, the Colton Formation is interfingering with the Green River Formation and the Flagstaff Formation is directly overlain by the Green River Formation.

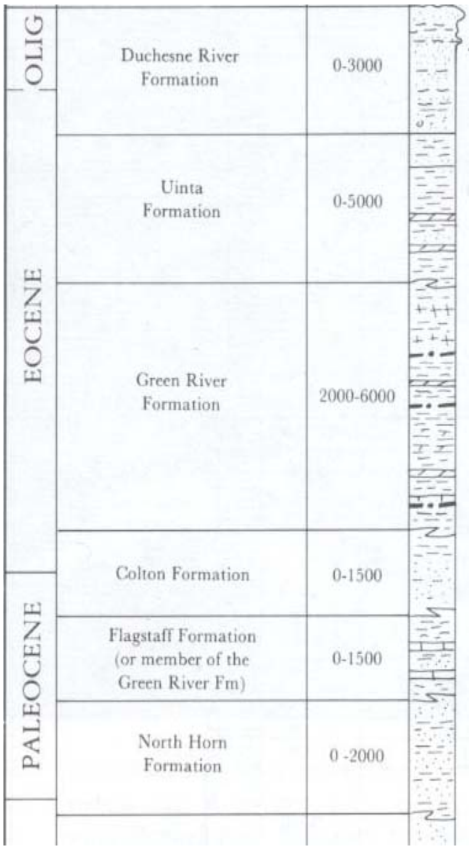
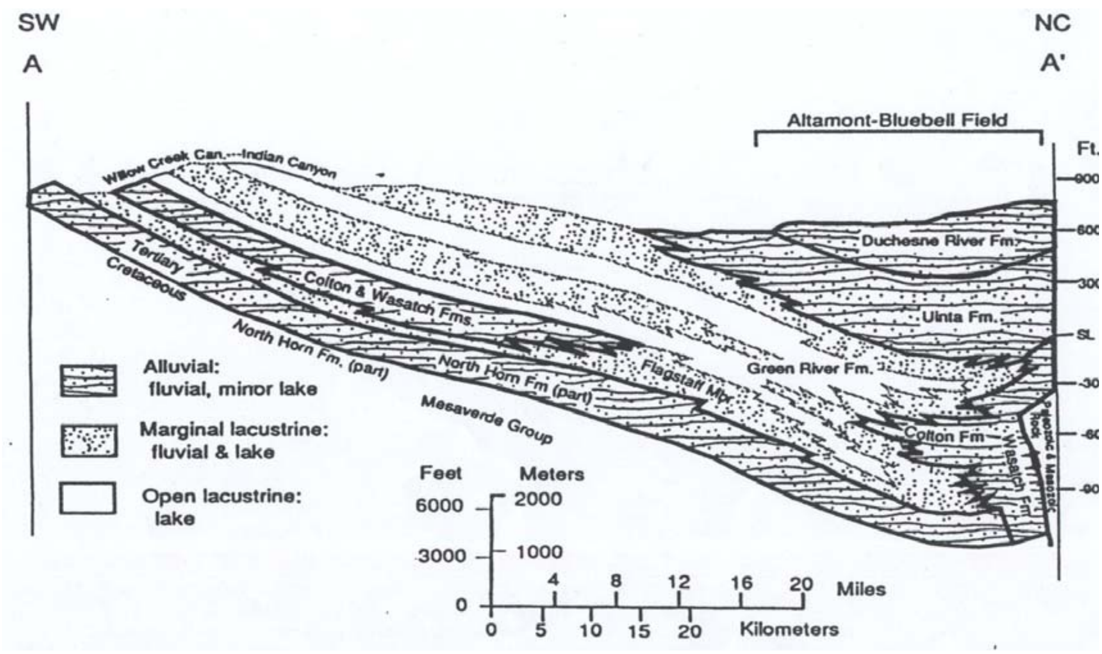


Figure 4.4 (left): Idealized stratigraphic column of the Uinta Basin (Hintze, 1988).

Figure 4.5 (below): Simplified stratigraphic transect of th Uinta Basin (Fouch et al, 1991)



To the west of Sunnyside, Flagstaff and Colton Formations are also interfingering, and east of Sunnyside, the Colton Formation is directly overlying the North Horn Formation. The combined North Horn-Colton Formation without the intervening Flagstaff Formation is in this area and in western Colorado also called the Wasatch Formation. The dating of the Colton Formation is debated; generally it is placed around the Paleocene/Eocene boundary. It does not cover the whole area of the Uinta Basin (Fig. 4.6). Stratigraphic thickness of the formation is over 1000 meters at Sunnyside (Fig 4.7) and decreases north-westwards (Fig. 4.6). The deposits were building out from the basin margins in the early Eocene and later retrieving to be overlain by lacustrine sediments. In the area of the present study, the Colton Formation forms a coarse-clastic alluvial wedge into the lacustrine Green River Formation, deposited from the south-east or east-south-east, pinching out to the west.

The Green River Formation is a lacustrine formation of green clay-, mud-, and limestone and for the most part overlying the North Horn Formation in a lateral and vertical facies transition. The Green River Formation is usually separated from the Colton Formation on the basis of sediment color, changing from lacustrine green/grey to subaerial red/brown, or on the basis of the lacustrine ostracode content. The sediments show that the Lake Uinta usually had alkaline water with high sulfur and carbonate concentrations. The sediments were thus quickly cemented, and the alkaline water is an indicator of a dry, hot climate. They indicate drier conditions than supported by the continental fossil record.

The Uinta Formation overlying the Green River Formation shows a regression and a last transition from lacustrine to fluvial facies in the area, indicating a termination of subsidence due to Laramide movements in the Eocene.

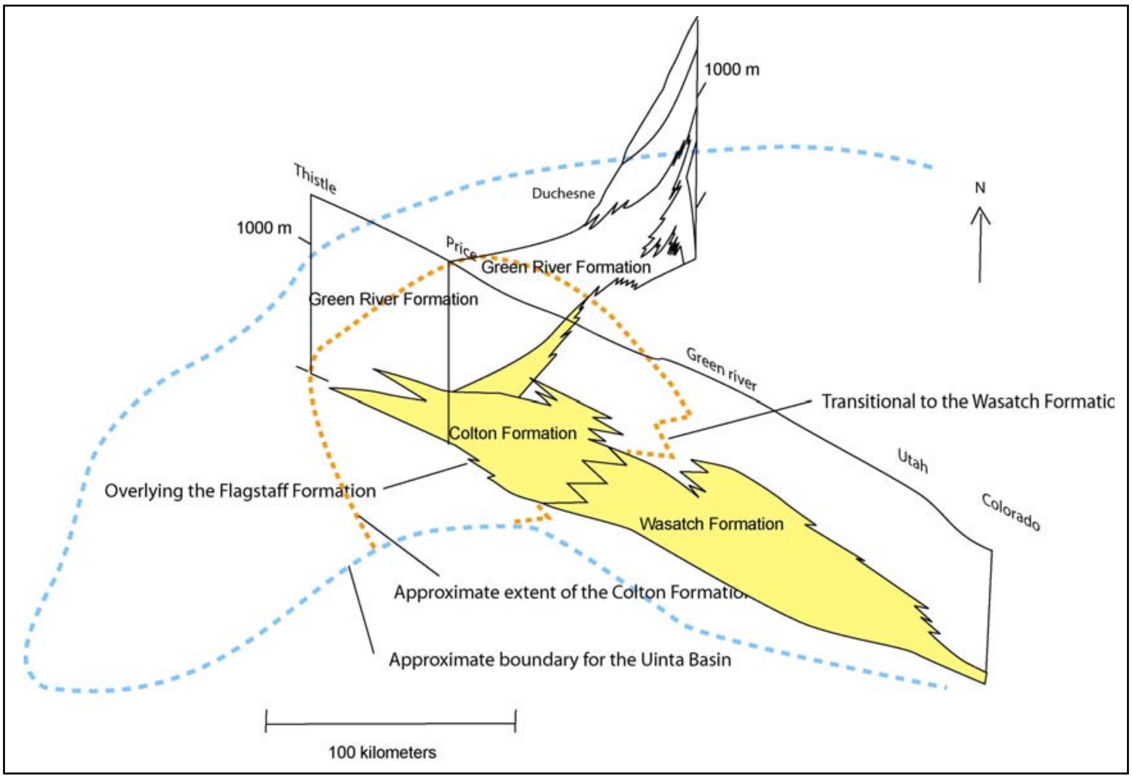


Figure 4.6: Stratigraphic sections of the Colton Formation in the Uinta Basin modified from Hintze (1988) and Fouch et al. (1991).

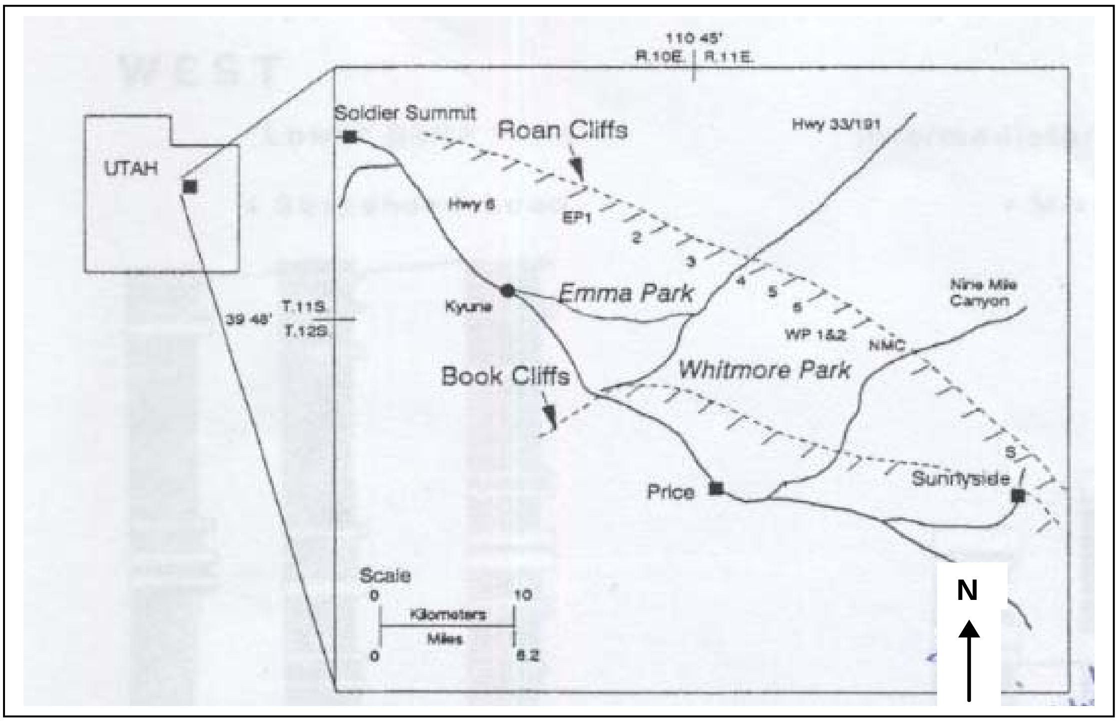


Figure 4.7: Area of extent for the Colton Formation (Morris et al., 1991)

4.3 Development of the Colton formation

The formation resulted from the progradation and later retrogradation of a coarse-clastic alluvial wedge from the east or southeast into shelf areas of the Lake Uinta. The Colton Formation shows a development from the east to the west from Whitmore Park to Solider Summit (Fig. 4.7), of sand-dominated fluvial channel and overbank deposits through mixed deposits to interdistributary bay, overbank, crevasse splay, mud flat and singlestorey distributary channel deposits. Near Sunnyside, in the eastern part of the Roan Cliffs, the sandstone percentage is about 70%, and thus a bedload system, in the Roan Cliffs at Whitmore Park it is about 50%, whereas at Emma Park still further west, the sandstone percentage is 10-20%, a suspended-load system (Morris et al, 1991). As well as being progressively more mud-dominated to the west, the formation also thins in this direction. The distance from Sunnyside to the distal part of the alluvial plain system in the west at Soldier Summit is 60-70 kilometers. From the study are in the Roan Cliffs in Whitmore Park to Soldier Summit the distance is 35 kilometers.

The sandstone deposits in the Colton Formation are classified as feldspathic wackes (Fouch et al., 1991) containing 25-75% quartzofeldspatic grains and usually calcite cemented (Taylor and Ritts, 2004) The probable provenance area is in the south and south-east, as Sevier-derived clasts probably would have been more coarse-grained and lithic (Fouch et al., 1991). Clast size ranges mostly from coarse-grained sand to mud; these sub-mature sediments may have been redeposited from alluvial plain sediments in front of the Monument and Uncompagre uplifts in the south and east (Stanley and Collinson, 1979). Large clasts are of intraformational origin, mudstone or pedogenic carbonate. Even though palaeoclimatic indicators in the fossil record does not support that the Paleocene/Eocene Uinta Basin area was subjected to large-scale droughts, the intraformational clasts, as mudstone chip lags, indicate at least prominent fluctuations in water discharge.

5 Facies

A sedimentary facies is a distinct lithological unit defined on the basis of color, bedding, composition, texture, fossils and sedimentary structures. It can be a bed or a bedset which is formed under certain sedimentary conditions, reflecting a specific process or environment (Reading, 1996). Facies definition and identification is based on purely descriptive criteria, but the facies are interpreted in a genetic context.

5.1 Table of facies

The main sedimentary facies are A: conglomerate, B: sandstone, C: siltstone and D: claystone, which are subdivided according to sedimentary structures as seen in Table 5.1 below. Their internal distribution is often described below in a context of *channel belt amalgamated sandstone* or *floodplain fines*. These terms are describing easily observable features in Colton Formation outcrops, as they are vertical and horizontal continuous intervals dominated by sand or silt-sized grains, respectively. Chapter 6, Facies Association, gives a detailed description of these features.

Table 1: Facies in the Colton Formation.

Lithology	**Facies**	**Structures**	**Interpretation**
Conglomerate	A1	Mud clasts	Dunes or lag
	A2	Mud- and calcrete clasts	Dunes, lag r slump
Sandstone	B1	Tabular cross stratified	2D dunes
	B2	Low angle trough cross stratified	3D dunes
	B3	Trough cross stratified	3D dunes
	B4	Planar stratified	Traction carpet, sheet bodies
	B5	Planar laminated	Traction or suspension, sheet bodies
	B6	Cross laminated, assymetric	Current ripples
	B7	Structureless	
Siltstone	C1	Cross laminated, assymetric	Current ripples
	C2	Planar laminated	Weak traction, suspension
	C3	Structureless	
Claystone	D	Structureless	Fallout from suspension

The proportion of interpreted facies by sections is shown below in Fig. 5.1. The general trend in the facies distribution, disregarding log 17-1, is that the structureless siltstone facies (C3) is dominating. Of the sandstone facies, low angle trough cross stratified (B2), trough cross stratified (B3) and structureless (B7) forms the most substantial part. In Fig. 5.1, 17-1 seems to be deviating from the other sections with respect to the percentage of sandstones present. This should be ascribed to 17-1 being merely a 40 meter long log taken from a small, protruding cliff at the eastern end of panel 17, at a level in the outcrop below the lowest point of the other logs. The cliff consists mainly of sandstone, which is most resistant to erosion.

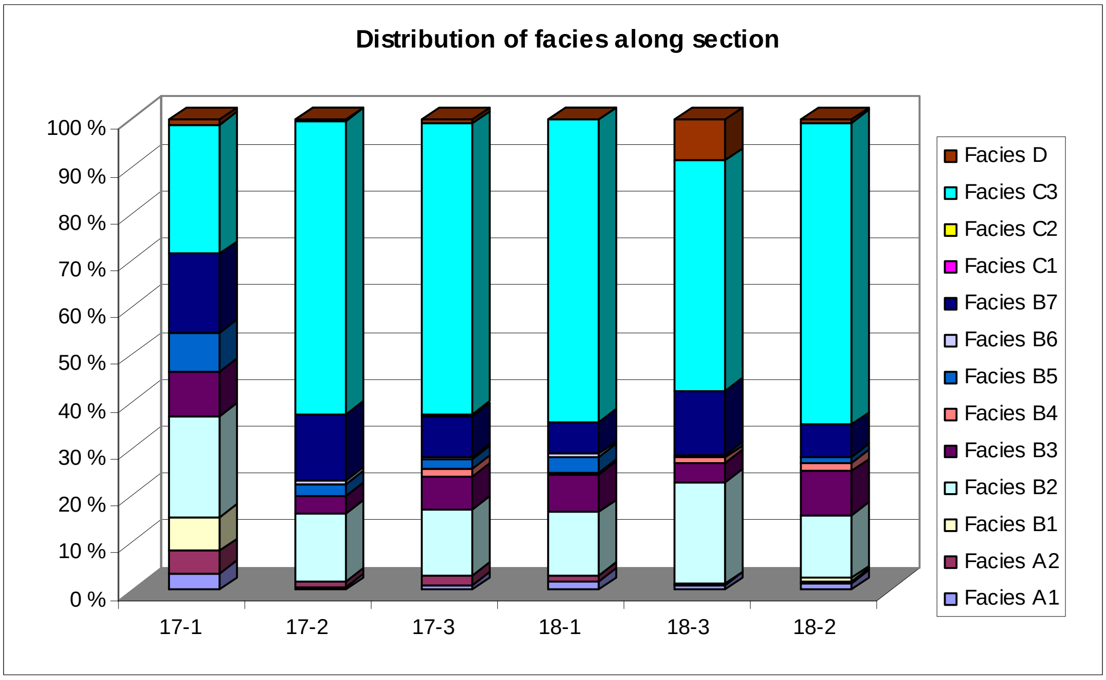


Figure 5.1: Percentage of facies in the logs along section east to west. Note higher percentage of sandstone and claystone in log 18-3. High claystone percentage can most likely be ascribed to good outcrop quality.

5.2 Facies A: Conglomerate

Facies A is a conglomerate facies that is subdivided in two main types: facies A1, which is a mudstone clast conglomerate and facies A2, which is a mixed mudstone- and calcrete clast conglomerate. Conglomerate beds are lenticular or tabular. Thickness of the conglomerate facies beds is varying, up to 3 meters, but usually less than 0.5 meter.

Facies A is often underlying or interfingering with beds of sandstone facies B, and almost always occurs above erosive bounding surfaces below, between and on top of clinothemes in channel sandstone bodies (Fig. 5.2). Sandstone clinothemes are distinct features of the channel belt amalgamated sandstones; laterally accreted successive lense-shaped bodies of sandstone with a moderate or low inclination. A further description of clinothemes is given in chapter 6. Tabular and lenticular conglomerate beds are of 2-20 meters length with graded lenses pinching out upwards on sandstone clinothemes. Conglomerate is also observed outside the channel belt sandstones, as the examples recorded in log 18-2 (208 meters) and 17-3 (98 meters).

When the lower boundaries of conglomerate facies A beds are lower channel boundaries they are usually strongly erosive surfaces. The erosion have a relief on a meter scale, up to 6 meters, cut down into amalgamated sandstone bodies or floodplain fines which may show paleosol development. Scouring is seen, with up to 1 meter relief. The upper boundary of conglomerate is erosive (0.1-0.2 meter erosive relief) or non-erosive surfaces below an overlying amalgamated sandstone facies. The color of the conglomerate facies is usually grey to yellow, but in 18-1 (295 meters) it has a green matrix.

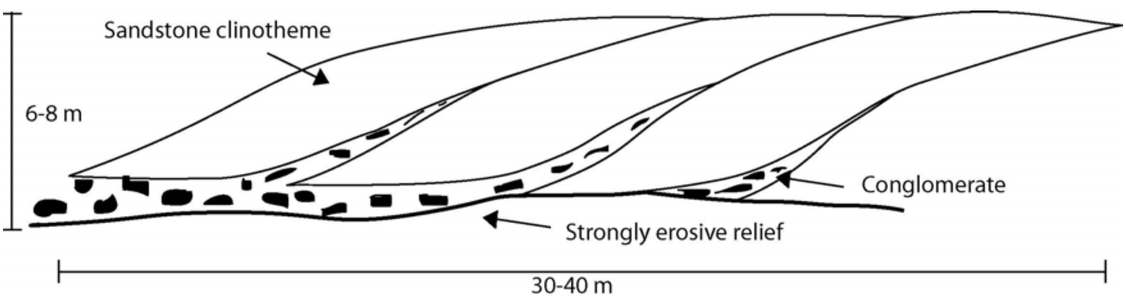


Figure 5.2: Typical conglomerate facies apperance.

5.2.1 Facies A1: Mudstone clast conglomerate

Description:

The facies A1 is a clast or matrix supported mud clast conglomerate found in thin beds or lags. Conglomerate facies A1 beds are found extending to the top of the sandstone clinothemes, but rarely as a basal lag conglomerate in channel units. Beds of mudstone clast conglomerate are usually bounded below by slightly erosive surfaces (0.1-0.2 meters) on the top of sandstone clinothemes or by strongly erosive surfaces towards siltstone or between amalgamated sandstone bodies. Matrix of this conglomerate is grey, yellow or red sand or silt.

Mudstone is used as a term of undifferentiated sedimentary rocks of grain size below silt, as applied by Boggs (2001). The color of the mudstone clasts is grey when unweathered, but usually appears yellow due to weathering. Clast shape is mostly tabular; sub-square, polygonal or rectangular seen in sections paralleled the longest axis. Clast size of the mudstone clasts is usually of pebble size, but can be up to 20 centimeters on the longest axes. The shortest axis is up to 5 centimeters, but commonly less. Small clasts are often more rounded than larger clasts, but most clasts are angular or only slightly rounded. The clasts are often oriented with the long axis parallel to the erosive surface.

Observed primary structures in this facies are low angle trough cross stratification and trough cross stratification with a foreset height of 0.3-0.6 meter, found in log 17-1 (22 meters) and 17-2 (256 meters), seen in figure 5.3. Sand lenses of up to 2 meters length and 0.3 meter height within the conglomerate are common, exemplified in log 18-3 (126 meters).

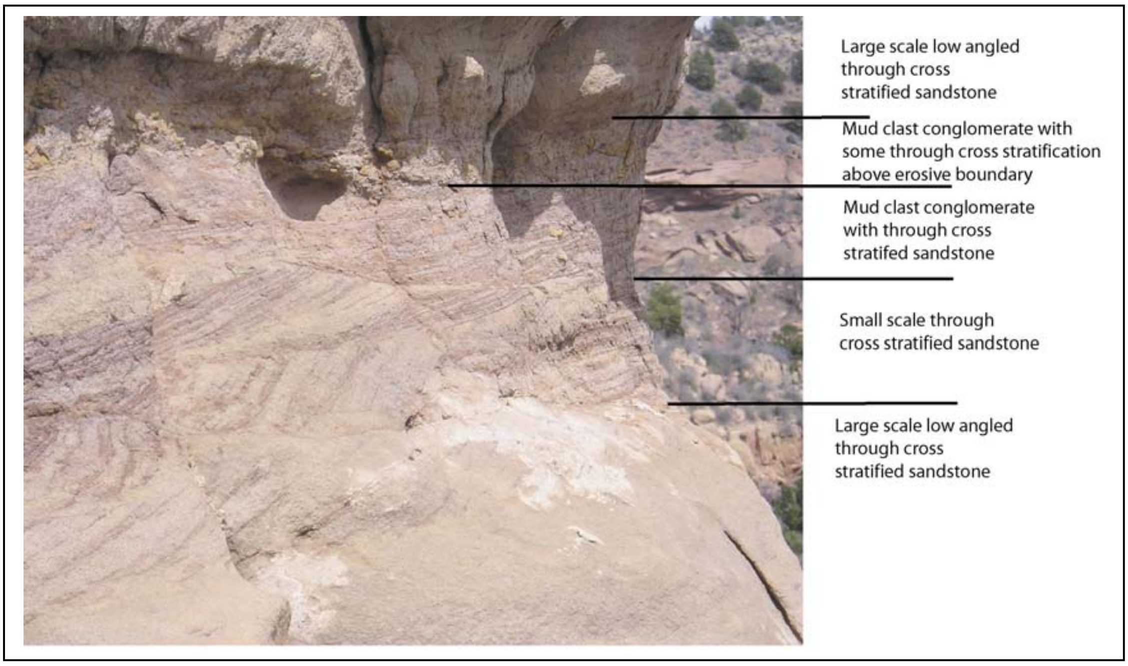


Figure 5.3: Facies A1 mud clast conglomerate from log 17-2, 256 meters. Height in left corner of the picture is 1 meter. The size of the uppermost mud clasts indicates that they may have accumulated through several fluctuations of water level in the river.

Figure 5.5: Thin bed and lense of facies A2 calcrete nodule/mud clast conglomerate, marked by arrows from log 18-3 (116 meters)

Interpretation:

Mudstone clast conglomerate is interpreted to have been deposited due to fluctuations in water discharge, affecting both water level and hydrodynamic regime. The fluctuations cause deposition of fines during falling water level and low flow regime (e.g. Allen, 2000). Increasing flow velocity and heightened water level will erode the semi-consolidated mud deposits and resediment the rip up mud clasts on the reactivated surfaces. The trough cross stratification structures indicates a relatively high hydraulic regime, with conglomerate being deposited as 3D-dunes, interpreted from Harms (1975) stability diagram for bedforms (figure 5.4)

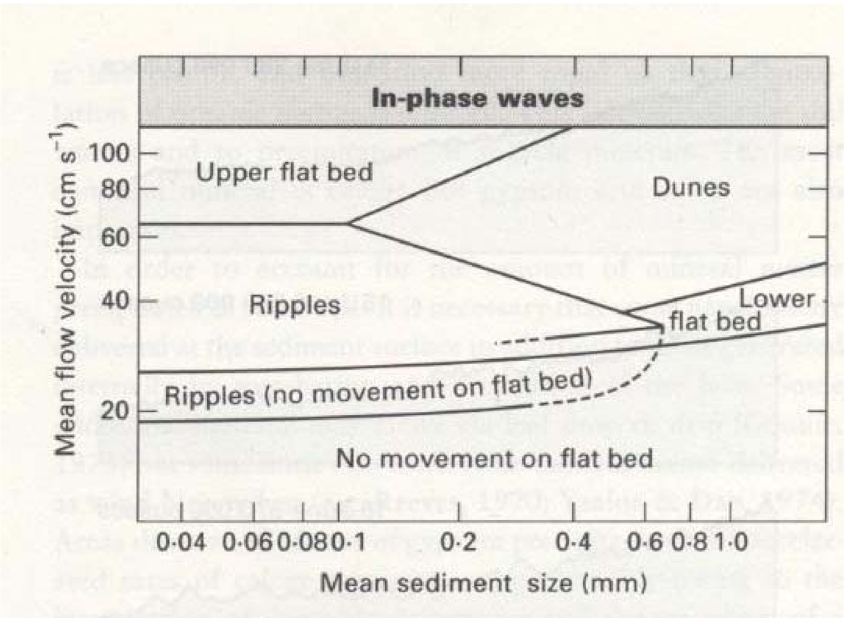


Figure 5.4: Equilibrium conditions for bedforms plotted by grain size and current velocity. Modified by Collinson (1996) from Harms (1975).

The largest mudstone clasts have been eroded from channel margins, and is actually similar to facies A2, but in this case there either are no calcrete clast development in the paleosol, or the calcrete nodules has been dissolved by water flowing through the sands postdepositionally. A likely example of the latter can be seen in 18-1 at 25-34 meters, where there is registered paleosol developed beneath the sand but no nodules in the conglomerate or lags. Carbonate concretions are found within these channel deposits, and may be precipitations of dissolved calcrete nodules.

Clast size may also be thought to reflect number of flood episodes and subsequent mudstone deposition during falling flow before final erosion and deposition of the lag. Several flood episodes probably in result thick mud deposits before a flood event is eventually large enough to erode the consolidated mud on the upper part of the clinothemes.

5.2.2 Facies A2: Calcrete nodule/mud clast conglomerate

Description:

Conglomerate bodies carrying calcrete nodules and mudstone clasts are lenticular and tabular. Bed and bedsets forming a basal lag in a channel sandbody often have wedges of conglomerate extending up from a basal lag as a lenticular body along sandstone clinothemes and pinching out upwards. When pinching out on clinothemes, the conglomerate is graded, being coarser grained in lower part than in the upper part. The mudstone clasts are both angular and slightly rounded and vary in size from less than one centimeter up to a few tens of centimeters Maximum size observed is 40 centimeters on the longest axis (log 18-1, at 170 meters). Large clasts are rarely rounded and some of the clasts are tabular. The calcrete clasts in the conglomerate facies consists of calcrete nodules, spherical or irregularly spherical. Clasts are up to 5 centimeters in diameter, but most clasts are 0.5-1 centimeter in diameter. In some beds the clasts in this conglomerate facies consist of only calcrete nodules, as seen in log 17-3 (3 – 8 meters into the log). Typical calcrete nodule/nud clast conglomerate is seen in figure 5.5)

Primary structures are tabular cross stratification with a foreset height of 0.1-1 meter (log 17-1, at 22 meters) and trough cross stratification. Very poorly sorted, angular mudstone and calcrete clast conglomerate beds showed a lateral grading away from a cut bank (log 18-1, at 170 meters). Facies A2 appear matrix supported in the most poorly sorted deposits.

Interpretation:

Tabular and trough cross stratified mudstone conglomerate beds are deposited as 2D- and 3D-gravel dunes respectively, maybe on a downstream accretionary bar or bank-attached bars. The calcrete nodules and the mudstone clasts are interpreted to be derived from paleosols eroded by channels (paleosol is described below in Secondary structures). Conglomerate with only calcrete clasts indicates that there has been reworking of the clast material by transport, which may have completely eroded the softer mudstone clasts. When conglomerate appears to be matrix supported, it might also be explained by that the paleosol is loosely consolidated outside the calcrete nodules, so that original mudstone clasts now appears to be matrix due to destruction through compaction.

Gravel sized clasts requires stream power corresponding to upper flow regime (Fig. 5.3) hydraulic conditions to get transported. Conglomerate may also be deposited as a result of bank collapse of consolidated or semi-consolidated sediments. These conglomerates are bank-attached, unsorted deposits with angular clasts (log 18-1, 170 meters).

5.3 Facies B: Sandstone

Facies B is a sandstone facies appearing in isolated beds, bedsets or packages of varying thickness and extent throughout the Colton Formation. Grain size is ranging from medium to very fine, of which fine is the most common grain size. Facies B is subdivided into facies B1, which is tabular cross stratified sandstone facies, facies B2 low angle trough cross stratified sandstone, facies B3 trough cross stratified sandstone, facies B4 and B5 are planar stratified and planar laminated sandstone, respectively, facies B6 is ripple-laminated sandstone and facies B7 is structureless sandstone.

In outcrop the facies B shows more resistance to erosion and weathering than lithologies dominated by silt or clay, and is commonly exposed as steep cliffs or protruding benches. The packages of sandstone facies can be enduring throughout the whole section laterally, forming continuous cliffs, but can also be of shorter extent and show a lateral transition to siltstone facies. Single beds or small bedsets can also have a large lateral extent, up to several hundred meters, or be deposited as small, thin lenses of a lateral extent ranging from 4 - 20 meters.

5.3.1 Facies B1: Tabular cross stratified sandstone

Description:
The tabular cross stratified sandstone facies B1 occurs as yellow to grey sandstone units deposited as single beds in larger packages in the amalgamated channel sandstones. The lateral extent of the facies varies between 5 – 20 meters, but can be difficult to estimate due to limited outcrop. In outcrops where the whole lateral extent was exposed, there was a lateral conform bounding or transition of facies to low angle trough cross stratification (B2).

Bed thickness ranges from 0.15 – 2.5 meters, only one bedset was observed (log 17-1 at 2 meters and 29 – 31 meters and Fig. 5.6). Foreset thickness is identical to bed thickness. The lower boundary is erosive, either planar or with an erosive relief of up to 0.2 meters on individual beds. The upper boundary can be gradational into ripple- laminated sandstone (B6) or erosive when facies B1 beds are overlain by units of the trough cross stratified sandstone facies (B2).

The tabular cross stratification is present both as high and low angled foreset units. Foresets generally change to be tangential towards the lower boundary. Sandstone sorting is moderately good, and the dominant grain size ranges between medium to fine sand. The beds can show grading.

Interpretation:

Tabular cross stratified sandstone represents 2D-dunes, which are straight crested dunes with parallel foresets when viewed perpendicular to the flow direction. 2D-dunes are deposited in the lower flow regime in flow velocities lower than the velocities required to form 3D-dunes provided the grain size is the same (Boggs, 2001). Sediments showing high angled foresets are deposited during lower flow velocities than low angled foresets with a tangential lower contact.

Facies B1 is deposited either as a channel floor downstream accretion bar or as dunes migrating across a larger bedform. Generally, simple bedforms as 2D-dunes are formed at deeper water than more complex bedforms as 3D-dunes or ripples (Boggs, 2001). Lateral transition to bedforms such as low angle trough cross stratified dunes is an indication of changing character of a migrating dune due to increased flow velocity (Fig 5.3). Vertical transition to rippled sandstone is most likely an indication of lower flow energy, or shallower depth. Since the bar does not have an erosive upper boundary, it might also be that the ripples are a preserved ripple train advancing over the bar.

5.3.2 Facies B2: Low angle trough cross stratified sandstone

Description:

The facies B2 consists of yellow to grey, low angle trough cross stratified sandstone. The facies can sometimes be found as single beds or small bedsets deposited in floodplain fines, but more commonly in beds or bedsets as part of the channel belt sandstone bodies. A significant part of the channel belts is of this facies.

The thickness of one bed is identical to the foreset thickness, ranging between 0.5 – 3 meters; a common value is 1.5 meter. Bedsets range in thickness from 0.5 – 5 meters, with values commonly between 2 - 4 meters. The lateral extent of beds and bedsets of B2 in the amalgamated sandstones in the channel belt can be up to 20 meters (Fig. 5.7), but beds and bedsets are often truncated below bedsets of the same facies. Single beds of sandstone within the floodplain fines may be several hundred meters, and are rarely truncated by erosion. However, there is often a horizontal facies transition to other sandstone facies in these beds.

Above siltstone or conglomerate the lower boundary is an erosive surface, with scouring cutting up to 1 meter into siltstone. Within multiple bedsets of low angle trough cross stratified sandstone both upper and lower boundaries are erosive surfaces. The erosional relief is usually less than 0.2 meter, but it can also be larger, truncating whole beds. Between low angle trough cross stratified sandstone and overlying trough cross stratified sandstone (B3) the boundary is very often transitional, in some cases also graded in the transitional zone. Boundaries below overlying trough cross stratified sandstone (B3) are also conform or moderately erosive on a scale up to a few tens of centimeters.

Primary structures found in this facies are the low angle cross trough stratification, sometimes with cross lamination superimposed on the troughs. Foresets are both straight and convex upwards. Grain size of this facies ranges from very fine to lower medium sand fraction, and the sorting is moderately good. When beds are graded, they are usually graded the upper 1-0.5 meter of a bed or bedset with a strong erosive upper (up to 2 meters) boundary. Bedsets can also show grading for as much as 4-5 meters. Inverse grading occurs in the lower 0.5 meter above an erosive boundary.

Slumping and water escape structures have distorted the primary structure in some intervals. This is most prominent in the upper part of the profile, where it also appears along with zones of patchy cementation, see i.e. log 17-3 (247 – 262 meters). Load casts are found above floodplain fines, seen in log 17-2 184 m.

Interpretation:

Low angle trough cross stratification is sections of 3D-dunes deposited in upper lower or lower upper flow regime. 3D dunes are a very common bedform in fluvial deposits, and it is also considered typical for beach deposits.

Thick beds are deposited when there is an abundance of clastic material in transport. In log positions 17-3 (258 – 262 meters) and 18-1 (282 – 290 meters) there is shown transitions from thick beds of low angle trough cross stratified sandstone to massive structureless and/or slumped sandstone. Bed forms with this internal organization are probably formed in rapidly deposited large amounts of material (Allen, 1984), which is often associated with flood.

Grading may indicate a reduction in flow velocity or shallowing, and the thickness of the graded bed or bedsets is dependent on the time span. Thus very thick continuously graded bedsets reflects a slower reduction in flow velocity than single beds. Inverse grading in the lower part of bedsets reflects an increase in flow velocity, as one could find in the early stages of a flood.

Load casts is generally attributed to dense sediment load upon less dense, usually loosely consolidated, sediments (Allen, 1984).

5.3.3 Facies B3: High angle trough cross stratified sandstone

Description:

This facies is a cross stratified sandstone (Fig. 5.8) deposited in bedsets of thicknesses between 0.6 – 9 meters. The most common values are found in the interval 0.6 – 2.5 meters. Color is yellow/grey, sometimes red/brown.

It is a characteristic facies within the channel belt sandstone bodies, but it also appears in single beds or smaller bedsets of sandstone in the thick packages of

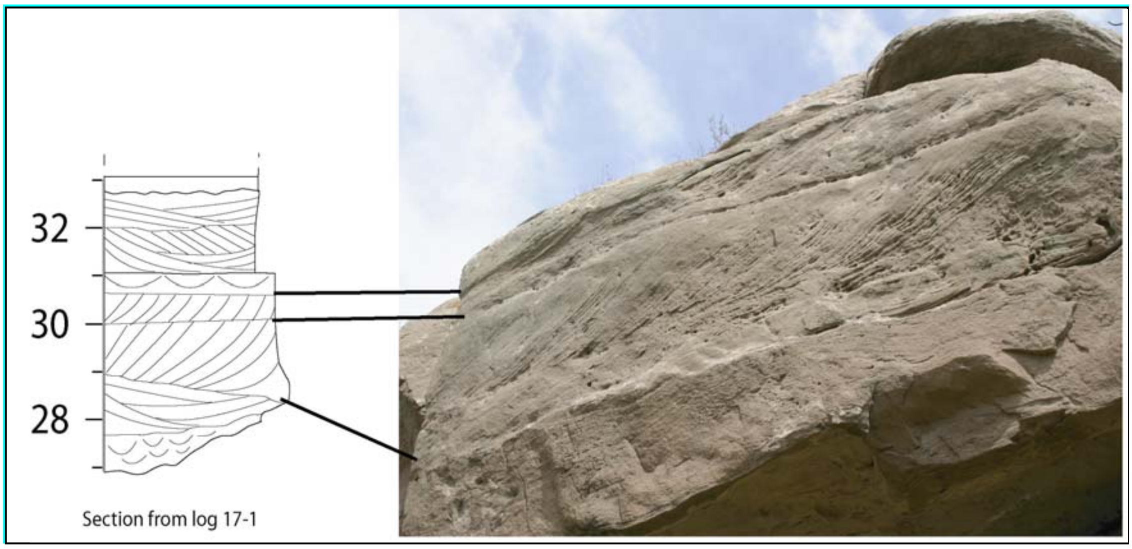


Figure 5.6: Facies B1 from log 17-1, 28 meters.


Figure 5.7: Facies B2 from the central channel belt. Length of dune approximately 10 meters


Figure 5.8: Facies B3 from log 18-1, height 32-34 meters into the log. Note carbonate concretions to the left.

floodplain fines. The beds consist of one or several foresets of thicknesses ranging between 0.1 – 3 meters. Although the bed thickness of all B3 beds ranges between 0.2 – 3 meters, the majority of the beds are within the 0.2 – 1 meter range (Fig. 5.9). The thin beds (approximately 0.2 -0.6 meter), most frequently occurs as single beds or small bedsets within the floodplain fines or directly above channel sandstone units. The 0.8 – 1 meter thick beds most typically occur in the channel belt sandstone units. An upwards transition from relatively large, intermediately angled trough cross stratification to smaller, high-angled troughs are commonly seen within the channel sandstone package.
Lateral extent of facies B3 beds and bed sets within the channel belt sandstone bodies is commonly around 20 meters, but the bed forms are often truncated and eroded and overlain or laterally replaced by beds of the same facies. B3 facies sandstone units commonly also have lateral extents of a few hundred meters with lateral transition to ripple facies within finer-grained sandstone or siltstone (facies B6 and C1). This lateral facies transition occurs within the floodplain fines, where B3 facies also can be found in 4-5 meter long isolated lenses with a maximum height of 0.5 meter pinching out to both sides.

When B3 is deposited as single beds or small bedsets within the floodplain fines, the lower boundary is moderately erosive (up to 0.3 meter, usually lower) and the upper boundary is usually conform to overlying siltstone, although the upper boundary surface sometimes have a relief of 5-10 centimeters. Thin beds of this facies can show an undulating appearance, and have an uneven upper boundary which probably reflects depositional morphological features instead of later erosion.

If units of this facies form the lower part of channel belt sandstone bodies along with conglomerates, the lower boundary is usually erosive into siltstone with an erosive relief ranging from 0.1 up to 2 meters, occasionally showing groove casts. Within single channel sandstone bodies beds of the B3 facies may have moderate (10-30 centimeters) to slightly erosive on the cm-scale, or conformable erosive lower boundaries. The upper boundary may be transitional, with the B3 sandstone facies passing gradually into rippled facies, be moderately erosive (up to 40 centimeters of relief) overlain by beds of same facies, or be deeply erosive with low angle cross stratified sandstone or conglomerate on top of the surface (up to 1 meter relief).

The grain size is very fine to medium, sorting is moderate. The beds may show normal grading, both normal and inverse grading within the same bed or be non-graded. Secondary structures are ripples on the troughs and soft sediment deformation structures. Some bioturbation is found in this facies. In the bedsets found in the floodplain fines, it is represented by both 0.8-2 meter long *Scolithos* structures and by 0.5 meter long branched *Scoyenia* structures within the bedsets. Bioturbation is described in secondary structures below.

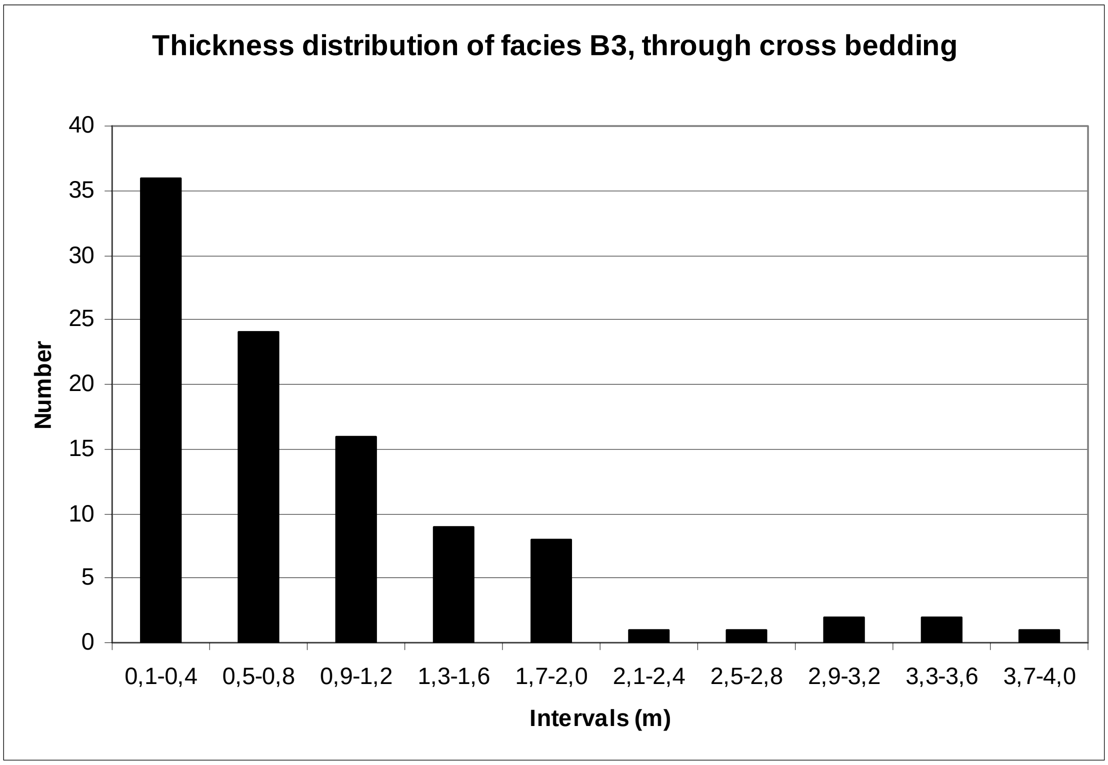


Figure 5.9: Thickness distribution of beds of facies B3.

Interpretation:

The facies B3, high angle cross stratified sandstone, represents 3D dunes deposited in upper lower flow regime at lower velocities or larger depths than low angled trough cross stratification. Upward decrease in foreset height may reflect decreasing depth or lower flow velocity.

5.3.4 Facies B4: Planar stratified sandstone

Description:

The B4 facies is a yellow or grey, sometimes red-brown planar stratified sandstone. Bed thickness varies between 0.6 – 3 meters, and the average thickness is 2 meters.

When the facies B4 occurs in the channel belt, it appears both as thin, 0.2-0.5 meter, and 1.5-2 meters thick sheet-like beds. Thin beds are conformly alternating with beds of low angle trough cross stratification in vertical successions (log 18-3, 60 meters). Thick beds of facies B4 are found above an erosional surface with a relief of 0.2 meters cut into structureless sandstone (B7). The upper conformable boundary is transitional and graded into climbing rippled siltstone (C1) or sandstone facies with low angle trough cross stratification (B2). Lower boundaries are conformable when the underlying beds consist of planar stratified sandstone or conglomerate (facies A), or the lower boundary is more transitional when underlying beds consists of cross stratified sandstone (B2-3) of same grain size. The stratification in this facies might have a small inclination, as seen in log 17-3 at 87 – 94 meters.

The datum chosen in the studied section is sandstone unit, consisting mainly of plane stratified sandstone, found in a 2-4 meter thick bedset of sandstones within the floodplain fines, with a lateral extent that can be followed throughout the panels 17 and 18 (except in log 18-1). Within the datum, planar laminated sandstone can be alternating with layers of low angle trough cross stratification (B3) and has an erosional lower boundary (0.2-0.4 meter erosive) towards trough cross stratified sandstone (B3) or low angle trough cross stratified sandstone B2). The lower boundary of the datum can also be B4 with an erosive boundary above floodplain fines and a thin layer of conglomerate with mudstone clast lining the erosional boundary of the datum in log18-2 (132 meters).

When planar laminated sandstone occurs in the floodplain fines, the sandstone deposits have upper conformable bounding surfaces towards overlying siltstone. The planar laminated sandstone is deposited both as single beds and bedsets with conformable lower bounding surfaces towards low or high angle trough cross stratified sandstone (B2 and B3). The bedsets have erosive lower bounding surfaces to floodplain fines with erosive relief up to 0.2 meter.

The grain size is very fine to fine with moderate sorting, mostly non-graded, but it can occasionally be normal graded.

Secondary structures registered are bioturbation represented by *Scoyenia* and *Scolithos*, with a degree of bioturbation classified as 1 -3. *Scolithos* burrow structures are up to 2 meters long, extending through and below beds and bed sets of this facies, mostly originated on the upper boundary of the B4 sand beds and within bedsets of B4. The width of the burrows is approximately 1 – 2 centimeters. The datum sandstone unit always has *Scolithos* structures, up to several meters long (Fig 5.10). *Scoyenia* structures are 0.2.-1.5 meter long and *Scoyenia* burrows have been recorded in log 18-2-2 and 17-3-2.

Interpretation:

Planar stratification is deposited as traction carpets in flow velocities above velocities required to form 3D-dunes (Harms, 1965 and 1982). Within the same flow velocities, planar lamination can be formed simultaneously as cross stratification on shallower depths in the channel (Harms et al, 1982). It can be typical of beach facies, being deposited by swash and backswash (Harms et al. 1982) or on alluvial plains in sediments deposited in one single event (Miall, 1996).

When thin beds of facies B4 conformly alternates with low angle cross stratified sandstone (B2) it reflects short-term fluctuations in flow velocity. An upward transition to climbing ripples (of facies B6) reflects a sharp decrease in flow velocity.

5.3.5 Facies B5: Planar laminated sandstone

Description:
The B5 facies is red-brown planar laminated sandstone, usually occurring within the channel belt sandstone bodies. Beds are either thin, 0.1-0.3 meter or about 1.5 meters thick.

Within the floodplain fines, planar laminated sandstone is deposited within bedsets of large lateral extent (several hundred meters) alternating with small scale (length of less than 0.5 meters and 10-15 centimeter foreset height) or low angle trough cross stratification (B3 and B2) and ripples (B6). This alternation occurs both as vertically alternating beds and as laterally alternating beds or as transitions between facies'. These bedsets have conformable or very slightly erosive (no more than 2 cm relief) internal bounding surfaces and conformable upper boundary to floodplain fines.

Within the channel belt sandstone units, 0.1-0.3 meter thick beds of planar laminated sandstone can be found alternating with beds of trough cross stratified sandstone (B2, also small scale) with conformable or slightly erosive boundaries (up to 10 cm erosional relief). Red beds of this facies also occur beneath major erosive boundaries forming the lower boundary below low angle cross stratified sandstone, with an erosional relief of up to several meters. The largest erosional boundaries cuts through beds above and continues into beds below the planar laminated sandstone bed. The beds are thus truncated within the channel belt unit and have limited lateral extension. The largest observed extent of the facies was 5 meters. Below the major erosive boundaries of the channel units B5 occurs as beds in graded bedsets showing a vertical transition from small scale trough cross lamination to B5 and ripples (B6) upwards. Vertical alternations of graded and non graded small scale trough and graded lamination in bedsets with conformable or very slightly erosive (up to 3 cm) surfaces also occur in the channel belt unit.

Bioturbation is represented by penetration of vertical *Scolithos* burrows with a bioturbation degree 2. The grain size is very fine to fine sand that is moderately sorted.

Interpretation:

Facies B5, planar laminated sandstone, can have been deposited from traction current in the upper lower flow regime or settled from suspension in the lower flow regime. The facies indicates a scarcity of input of clastic material when it has been deposited by traction in the upper flow regime. Alternation of current ripples and planar beds indicates deposition in very shallow water, or that the laminated beds were deposited from suspension by water splashing up on and over levees or scroll bars during flood. Depositional environments can have been channels, crevasse splays or beach facies. Low energy planar beds can be deposited in floodplain, oxbow lakes or levee. Beach facies planar laminated sandstone is a result of swash and backwash in the beach zone with scarcer clastic input than when planar laminated sandstone forms.

5.3.6 Facies B6: Cross laminated sandstone

Description:

The facies B6 is red/brown, occasionally yellow, current ripple laminated sandstone. Beds of the facies have asymmetric ripples which also can be climbing, as seen in log 18-1 at 179 meters and 18-2 at 85 meters height (Fig 5.11). Facies B6 consists of beds ranging between 0.2 – 1.7 meters, commonly around 0.5 meter. The beds can be single or comprise bedsets up to 4 meters thick (seen in log 17-2, at 119-125 meters), but usually the bedsets are about 1 - 2 meters thick.

When beds and bedsets of this facies occur within the channel belt sandstone units, their lower boundaries are usually conformable towards underlying low angle trough cross stratified (B2) and planar stratified sandstone (B4). The lower boundary to the other beds of facies B6 is in some places erosive surfaces with an erosive relief up to 0.5 meter. Both the upper and the lower boundaries can also be transitional relative to structureless (B7) or trough cross stratified sandstone (B2). The upper boundaries of beds of this facies in the channel belt unit are conformable to other beds of facies B6, with thin layers of siltstone or claystone and small scale trough cross stratified sandstone and planar laminated sandstone in the transitional border zone. Beds of current ripple laminated sandstone can also be transitional bounded into laminated sandstone (B5) with decreasing grain size in the transitional zone. When B6 is overlain by conglomerate (A), the upper boundary is an erosional surface.

Beds of facies B6 that occur within the floodplain fines have both upper and lower conformable or transitional bounding surfaces to sandstone facies. If the current rippled sandstone beds are succeeded by siltstone, the upper boundary is usually conformable, but can also be an apparently erosive surface. The lower boundaries of the sandstone beds are conformable to underlying siltstone or erosive surfaces with a relief up to 10 centimeters.

The grain size varies from very fine to medium, with moderate sorting. The beds are both inverse and normally graded, but also non-graded. Secondary structures are bioturbation represented by S*colithos,* classified as 3-4.

Interpretation:

Sandstone with asymmetric ripples is deposited from traction by unidirectional flow in the lower flow regime. Ripples migrate in the direction of the net current. Climbing ripples are deposited due to a rapid decrease in flow velocity and/or large amounts of sediments in transport. Transitional boundaries to other facies´ and grading (both normal and inverse) indicate gradual fluctuations in the hydraulic regime during deposition.

5.3.7 Facies B7: Structureless sandstone

Description:

Facies B7, structureless sandstone, is commonly red, but sometimes yellow or grey. It is characterized by scarce or no visible sedimentary structures. The facies is present in beds with thickness that ranges between 0.1 and 9 meters, but the majority of beds have thickness between 0.2 and 1 meter. In many cases the beds are isolated or divided by erosive boundaries*.* Thin beds of thicknesses 0.2-0.5 meter may form bedsets of thicknesses between 1 – 4.5 meters within the floodplain fines. The lateral extent of beds and bed sets of this facies is large. Sandstone units within the floodplain fines may be up to several hundred metersin lateral extent. Whether this facies actually is continuous within the sandstone units is difficult to establish. Closer inspection could very well show lateral transitions to other facies within the bedsets. The lateral extent can also be large within the channel sandstone belts, up to 50-70 meters or more, but lateral truncation by erosion is common.

The lower boundaries are generally conformable or slightly erosive (with less than 10 centimeters erosional relief) towards other facies; however, locally the lower boundary can also be stronger erosive. When under- or overlying beds consist of low angle trough cross stratified (B2) or structureless sandstone (B7) facies in the channel belt sandstone, the bounding surface can be erosive with a relief up to 2.5 meters, but more often less than 1 meter. Beds of this facies have also in some places been eroded up to 4 meters prior to deposition of overlying channel bodies with low angle trough cross stratified sandstone or conglomerate beds. The upper boundary is conformable or slightly and moderately erosive (up to 0.5 meters relief) when beds of structureless sandstone are overlain by other beds of structureless sandstone, silt and claystone. The upper boundary can also be transitional into small scale trough cross stratified or rippled sandstone.

Primary structures are at some places slightly visible, as ripples and low angle trough cross stratification. Soft sediment deformation structures can be seen in slightly visible layering (probably low angle trough cross stratification) in otherwise mainly structureless sandstone. Within the channel belt sandstone bodies, beds of this facies thinner than 2 meters can be sharply graded, often in combination with red coloring. Beds and bedsets within the floodplain fines shows both grading and inverse grading.

Grain size is ranging either between very fine/fine sand or fine/medium sand of moderate sorting. Secondary structures are bioturbation, both by vegetation and organisms. Bioturbation can be significant. Some bioturbation is found as *Scoyenia* structures up to 0.5 meter long, but mostly the type of bioturbation is not identified. On our scale from 1-6, where 6 is largest, the degree of bioturbation commonly lies between 4 and 5.

There are probably several explanations for the structureless deposits. They may initially be structureless; this may be due to rapid deposition as rapid fallout from

Figure 5.10: Facies B5 in datum, planar stratified sandstone facies with over two meter long traces of *Scolithos* bioturbation in negative relief.

Figure 5.11: Facies B6 with climbing ripples.

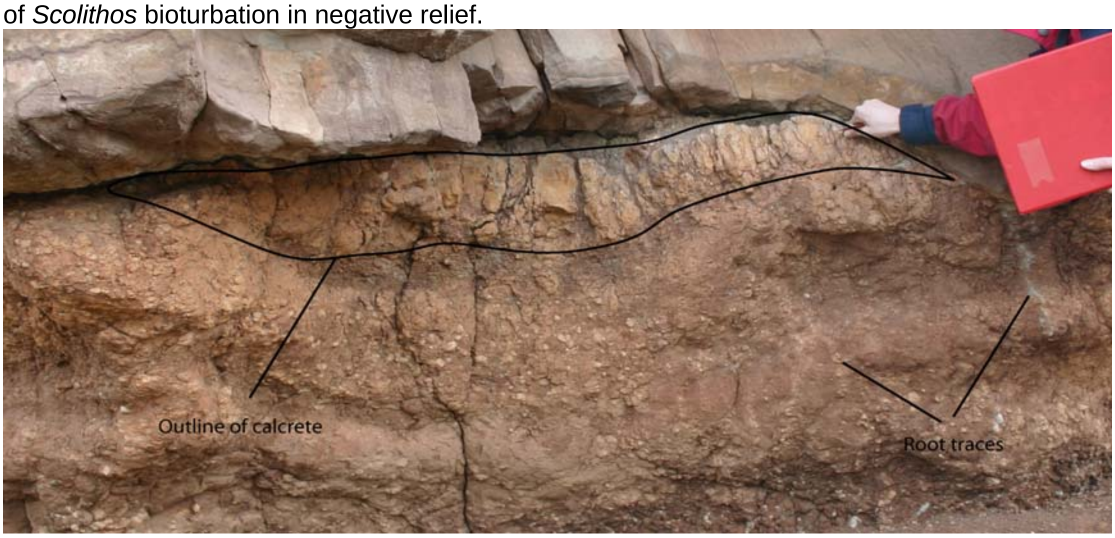


Figure 5.12: Paleosol with continuous calcrete layer eroded by facies B2 from

Interpretation:

suspension, as slumps and debris flows from unconsolidated channel banks due to strong floods (Martin and Turner, 1998).

Secondary structures may also have destroyed the primary ones, which is a possible explanation when the structureless sandstones are strongly bioturbated or deformed by soft sediment deformation. Early post-depositional dewatering and sediment movement is a likely cause when structureless sandstone is found in continuous deposits with the slumped trough cross bedded sandstone mentioned above (Collinson and Thompson, 1982). Bioturbation may have been especially significant in thin sandstone beds deposited within the floodplain fines. Some of these beds were found to be so intensely bioturbated that they could not be assigned to any ichtnofacies.

A last factor that has to be taken into consideration is that most of the outcrop along the Roan Cliffs is eroded and weathered, especially towards the top of the section, so existing structures may not be visible by this reason.

5.4 Facies C: Siltstone

Facies C is siltstone, and the most frequently occurring facies in the Colton Formation. It is subdivided into facies C1, cross laminated siltstone, facies C2 which is laminated siltstone and C3 which is structureless siltstone. Siltstone outcrops are easily eroded, and is commonly occurring as slopes of loose sediments in between the sandstone cliffs. The exposures are thus usually bad, and structures not seen for most part of this facies. Generally, the siltstone has high clay content, and it can in some cases be argued that parts of what is interpreted as siltstone is actually claystone. An alternative is to apply the previously mentioned common term mudstone. Floodplain fines, which are mainly characterized as facies C3, constitute in average 59 % of the total logged outcrop.

5.4.1 Facies C1: Cross laminated siltstone

Description:
Facies C1 is an assymetrically cross laminated siltstone which is observed within the channel belt sandstone body as deeply red colored beds. The facies C1 occurs in single, 10 centimeters thick layers alternating with planar laminated

siltstone and clay in a 1.5 meters thick package, or within structureless sandstone. Lateral extension is recorded to be 20 meters, and upper and lower boundaries are transitional/conform within siltstone and clay, and graded when bounded by sandstone facies. Example of this facies is shown in log 17-3-5.

Interpretation:

Assymetrically cross laminated siltstone is deposited as rippled beds in unidirectional lower flow regime current in shallow water. This facies can have been deposited in a falling stage of flood on levees or in floodplain sediments as distal crevasses. Cross laminated siltstone structures are not likely to be preserved or observed in floodplain sediments due to bioturbation or bad exposure.

5.4.2 Facies C2: Laminated siltstone

Description:

The C2 facies is a red/brown siltstone occurring in beds with thicknesses up to 0.3 meter. Beds of the facies have a limited lateral extent, less than 20 meters. Laminated siltstone drapes reactivation surfaces bounding high angle trough cross stratified sandstone units, or the siltstone beds are conformable and transitional into overlying small scale trough cross stratified sandstone (height 10-15 centimeters, length 0.5 meter). The upper boundary is erosive when cut by a channel surface, commonly with a relief exceeding 1 meter, thus truncating the laminated siltstone bed. Facies C2 can be found alternating with cross laminated siltstone (facies C2) in bedsets of a total thickness of 1.5 meters. Boundaries between the alternating beds may have thin clay laminae. Examples of this can be seen in 17-3-5 and 18-1-3.

Interpretation:

Laminated siltstone is deposited by traction currents or as fallout from suspension during falling flow velocity (Miall, 1996). In one locality (log 17-3, 254 meters) laminated siltstone is interbedded with claystone and asymmetric cross laminated siltstone. This indicates that the sediments have been deposited in very shallow water in the lower part of the lower flow regime, but in pulses under varying hydrodynamic conditions.

5.4.3 Facies C3: Structureless siltstone

Description:
The C3 facies, structureless siltstone, is the most abundant single facies in the Colton Formation. It occurs as beds and bedsets, but individual beds are difficult to identify. The lateral extent of depositional units of the facies is considerable, often continuous throughout the section, but the siltstone units can also be truncated by erosion beneath channel belt units or be thinning along the section.

The structureless siltstone bedsets are very thick, between 5 - 80 meters of seemingly continuous depositional units. The siltstone units are interrupted by thin beds or small bedsets of sandstone, structureless, planar stratified (B4) or small scale trough cross stratified (B3) and ripple-laminated (B6). Directly below these sandstone units there are visible paleosol development in the siltstone facies in some good exposures.

The structureless siltstone is dominantly red colored. Occasionally there are green beds visible in between the structurally undifferentiated red siltstone beds or laminae. These green beds are estimated to be on a scale of centimeters to a few tens of centimeters of thickness.

Units of structureless siltstone have lower conformable boundaries to all sandstone facies. Upper boundary can be conformable or erosive surfaces below sandstone units within the floodplain fines. When beds of the siltstone facies are underlying the thick, continuous channel belt sandstone units, the upper boundary is a strongly erosive surface, with up to 6 meter erosive relief.

There are few visible primary structures, most likely due to the high degree of weathering and erosion in rocks of this facies. The grain size distribution can be characterized as poorly sorted clay rich siltstone. There is inverse grading, or upward coarsening, into very fine sand seen in well exposed outcrops protected under sandstone cliffs (log 17-2, at 98 meters); however, there are also graded transitions to claystone (log 17-2 at 267 meters and 17-3 at 288 meters height).

Secondary structures are paleosol development with mottling and calcrete concretions, also especially visible directly below the channel belt sandstone bodies. Beneath channel belt sandstone erosional boundary in log 17-3 at 127 meters there is a zone of 0.8 meter where the color of this facies is grey, following the contour of the channel body.

Interpretation:

Clay rich silt facies that is apparently structureless can be deposited by rapid fallout from suspension on a floodplain due to falling water discharge or flow velocity. As discussed above in the interpretation of the structureless sandstone facies, there may be several reasons as to why deposits appear structureless. A reason for the structureless appearance of the siltstone facies may simply be the overall poor quality of the exposures, since most of them is covered by loose talus sediments. The red color of the siltstone facies is due to oxidation, green layers has been reduced due to reducing pore water. Green, thin layers may have been deposited in reducing conditions in shallow lakes on the floodplain, but may also have been reduced postdepositionally due to water-logged conditions (Collinson, 1996), or meteoric water circulating in fluvial channels after burial. Reducing pore water indicates a high organic content in the sediments, opposing the arid conditions indicated by calcrete nodules. More significantly: green fine grained sediments are also very diagnostic of the Green River lacustrine deposits, and green layers would then be a result of a lacustrine transgression.

5.5 Facies D: Claystone

Description:

The D facies, very fine grained sediment dominated by the clay fraction, is not often observed in Colton. The color varies from grey to reddish brown. Some of the thin, green beds within the floodplain fines are also claystone. Claystone occurs usually as thin layers in connection with sandstone bodies. The facies is found as drapes on beds and bedforms, or under and between channel sandstone bodies. Thin laminae is also seen alternating with laminated/stratified and rippled silt- or sandstone The layers are in a scale of centimeters to tens of centimeters (max 40 cm), and show little structures.. The upper boundaries of claystone beds are deeply erosive surfaces below overlying low angle trough sandstone (B2), and the bounding surface may show loading and scouring. The upper boundary may also be gradual or conformable towards thin layers of rippled sandstone or laminated silt. Lower boundary is conformable or graded above structureless siltstone*.* In logs 17-3-6 and 18-2-2 the siltstone package of floodplain fines shows a gradual transition to claystone.

Lateral extent of units of clay facies has been observed up to 50 meters, but the claystone units are often eroded and discontinuous. Deep erosion combined with relatively thin layers result in discontinuous layers of claystone within the channel belt sandstone package.

Interpretation:

Mud is deposited as fallout from suspension, and requires a very low flow velocity to be deposited. Colors of claystone reflect the same depositional environment as mentioned in the silt facies. Mud may have been deposited as clay drapes on top of channel levees, in standing water pools during channel abandonment or water low stand (Miall, 1996), or be deposited from suspension in a lacustrine setting. Clay-rich mud drapes sandy bedforms, but is more prominent within siltstone packages.

In association with sandstone bed successions, claystone units reflect the lowest energy of fluctuating flow regimes. In the silt package claystone beds may be the most distal parts of crevasses, thus being deposited simultaneously with a high hydraulic energy regime in the channel. When clay has been deposited upon an erosive surface, the erosion was clearly not done by the same process that deposited the clay. The mud draped erosive surface thus represents a break in sedimentation. This break might, however, not necessarily have been long, merely a result of seasonal changes.

5.5 Secondary structures

5.5.1 *Paleosol development*

Paleoped development is seen by red coloring, mottling, root structures and calcrete nodules and desiccation cracks. Dessication cracks are not common. Leached horizons within the paleopeds are found in good exposures below lower boundary of channel sandstone bodies. Paleosol maturity is ranging from 40 centimeters of mottling to up to 5 meter thick packages of calcrete nodule development within floodplain fines. Calcrete nodules were also seen within the loose sediments of weathered floodplain fines slopes, but extent was not established.

Calcrete nodules in paleosol are often increasing in numbers to the top of the paleosol, and the most mature development of calcrete found was a 10-15 centimeters thick calcrete layer with a blocky fracturing pattern on the top (Fig. 5.12). Their upper boundaries are channel floor or channel belt floor surface and lower boundary is transitional into floodplain fines or not well exposed. Calcrete indicate semi-arid or arid conditions, since calcrete is formed due to evaporation and capillary groundwater flow (e.g. Miall, 1996). Development of calcrete requires a considerable time span for full paleosol maturation (in the range of ten thousand years according to Retallack, 1988); dependent on temperature and mineral grain size, but the most important factor is low sediment input.

Root structures seen in the Colton Formation are up to 30 - 40 centimeter long, 0.5 centimeter wide risoliths or up to 1 centimeter wide mottling. The observed mottling and rizoliths are traces of vertically penetrating roots with a few downwards stretching branches, and the observed root systems would be expected to be found in settings with modestly sized vegetation (Retallack, 1988).

Wash-out zones or paleogley in the floodplain fines is an indicator of water-logged conditions (Boggs, 2001). Water-logged conditions can be very confined, limited to the lowest lying parts of the floodplain, where small, shallow lakes may form (Kraus and Brown, 1999). Wash-out zones in sediments beneath fluvial channel segments can be formed as a result of the channel sands and the sand-rich floodplain sediments under being an aquifer after burial.

5.5.2 *Bioturbation*

Bioturbation is interpreted to be represented by non-marine *Scolithos* and *Scoyenya* ichtnofacies where identified. Most of the bioturbation in the outcrop is not identified to any ichnofacies.

Scoyenia ichnofacies is characterized in Colton by up to 0.5 meter long vertical burrows. Burrows might be branched or horizontal to curved. Environment requirements for *Scoyenia* are freshwater or terrestrial firmground (Bromley, 1996).

Scolithos ichtnofacies is non-marine in this setting. *Scolithos* trace fossils is seen as long, vertical tubes (both in positive and negative relief) from 0.5 meters to several meters long and 1-2 centimeters thick. Along with the *Scolithos* tubes, there are vertical digging traces, up to 4 centimeters wide, most likely made by

Arthropods, as seen in Fig. 5.13. This ichnofacies is very typical of sandy high energy environments, as beaches, fan deltas or subaquatic crevasse fans ((Bromley, 1996; Jenø Nagy, personal communication). Trace fossils are very rarely found in a channel setting.

6 Facies associations

6.1 Table of facies associations and surfaces

Facies associations are assemblages of spatially and genetically related facies, distinguished as the principal "building blocks" of the sedimentary succession (Collinson, 1996). Interrelationship and grouping of adjacent facies is crucial for interpreting depositional environment and succession, as the facies themselves are ambiguous (e.g. Miall, 1977).

Below (Table 6.1) is a table of interpreted facies associations; Channel infill (I), overbank (II), floodplain (III) and Lake (IV). The percentage of these associations from the logs along the outcrop can be seen in figure 6.1. When Fig 6.1 is compared to figure 5.1, it is found that the variation is larger along section when facies are grouped into facies associations.

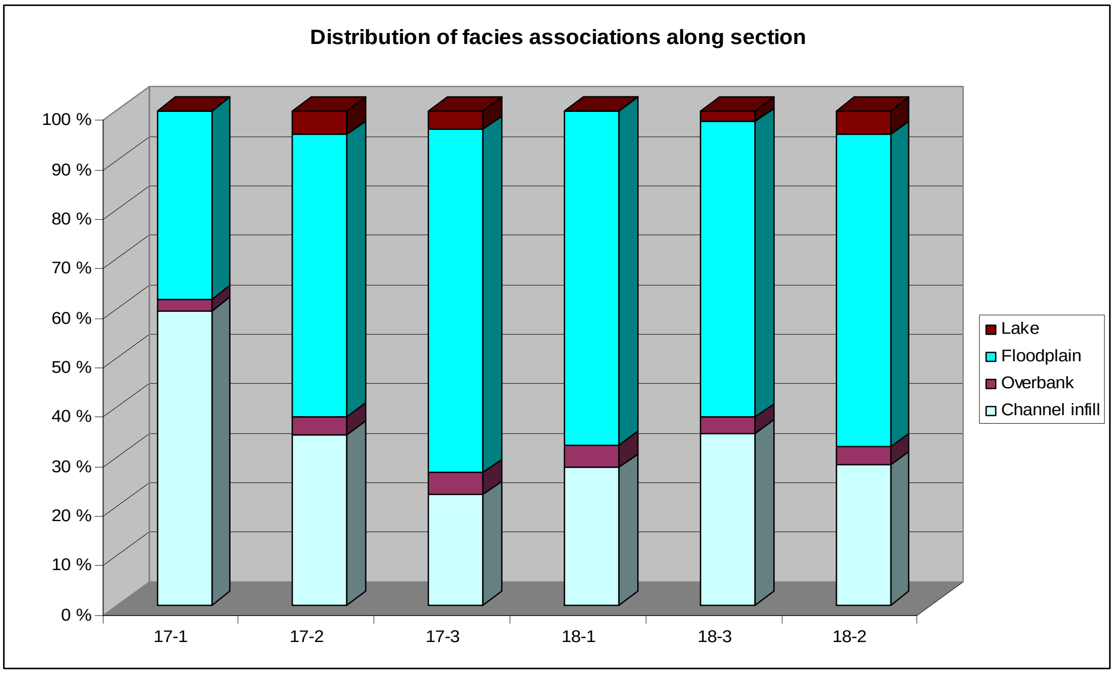


Figure 6.1: Histogram of percentages of facies associations in panel. Note high channel infill facies association percentage in 17-2 and 18-3. 17-2 is again deviating and should be ignored.

Table 6.1: Facies associations in the Colton Formation

Facies association	Interpretation of facies association	Facies grouping in association	Interpretation of sub-facies association
IV	Lake	D	Lacustrine claystone
		B3 B4	Beach deposit
III	Floodplain	B3 B4 B5 B6 B7	Crevasse
		C3 D	Mudstone
		C3	Paleosol
II	Overbank	B3 B4 B5 B6 B7 C1 C2 C3	Levée
		B2 B3	Crevasse channel
I	Channel infill	A2 A1	Channel lag
		A1 A2 B2 B3 B4 B6 B7	Lateral accretionary element
		B1	Downstream accretionary element

6.2 Facies association 1: Channel infill

6.2.1 Architectural elements in channel infill

A sedimentary architectural element is characterized by a facies assemblage, internal geometry, external form, bounding surfaces and, in some instances, vertical profile (Miall, 1996). In a hierarchy of depositional structures architectural elements are placed below the facies associations and above the single facies. Architectural elements represent smaller depositional events within a larger unit, as bars in a channel infill. In the channel infill facies association, the identified lower hierarchical architectural elements are lateral accretionary bars/point bars, downstream accretionary bars, braided bars/sandflat and abandoned channel infill.

The depositional setting of the architectural elements present in Facies association I, Channel infill, is illustrated in figure 6.2. Description and interpretation of the channel infill sandbodies are given below in 6.2.6.

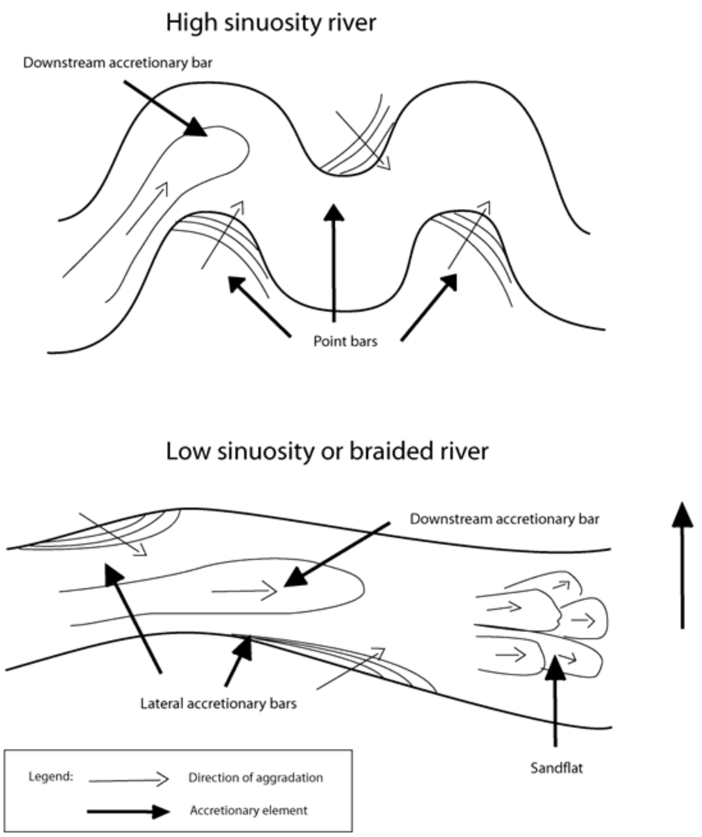


Figure 6.2: Areas of lateral and downstream accretion in channels. Direction of lateral aggradation of elements can show a large variability, and the direction given in the figure is only valid for the given bars.

6.2.2 Lateral accretionary bars/point bars

Description

Laterally accreted bars, which also include point bars, consist of mudstone clast and calcrete nodule/mud clast conglomerate facies (A1 and A2), tabular cross stratified sandstone (B1), low angle through cross stratified sandstone (B2), through cross stratified sandstone (B3), planar stratified sandstone (B4), cross laminated sandstone (B6) and structureless sandstone (B7) facies'.

Lateral accretion units (LA) are inclined sandstone lenses with dipping bounding surfaces (clinothemes), often with conglomerate drapes at the toe and occasionally mud clast conclomerate at the top. A fully preserved clinotheme has sigmoidal upper and lower boundaries, as shown below in figure 6.3. Clinothemes within a laterally accreted succession are bounded by reactivation surfaces. Erosional relief on the clinothemes varies, sometimes parts of the clinothemes are completely eroded and remnants of mud drapes are almost always only found as mud chip lags above erosive surfaces (figure 6.4).

Height of clinothemes varies between 2 to 14 meters; the average value is 8 meters. Length varies from 20 to 150 meters, and 50 is a common value. Lateral thickness of the central part varies from 1-2 meters to over 10, as is demonstrated in figure 6.3 and 6.4.

Bedform configuration within the lateral accretion elements varies both laterally and vertically, as can be seen in figure 6.4. A typical vertical facies development within a clinotheme is starting with conglomerate facies A1 or A2 at the bottom, sandstone facies B2 and B3 as large 3D dunes in the lower parts with internal slightly erosional or bedform (table 6.2) boundaries, changing upwards into small scale sandstone facies B3 and B6 towards the top. Also, tabular cross stratified sandstone (B1) dunes are occasionally found in the lower parts of clinothemes or smaller dunes within the clinothemes, and planar stratified sandstone (B4) and cross laminated sandstone (B6) towards the top. Scale of structures is often upwards decreasing. Vertical successions of grain size from the bottom upwards through LA units often display an upward fining development. Low in the central channel belt, however, the grain size seems to be rather stable through the vertical succession.

The top of the clinothemes may have a transition to levée elements; a graded transitional zone with conform surfaces. Boundaries between lateral accretion elements and crevasse, floodplain fines and some of the levées are sharp conform surfaces. The clinothemes are red colored towards the top where they show a transition into levées.

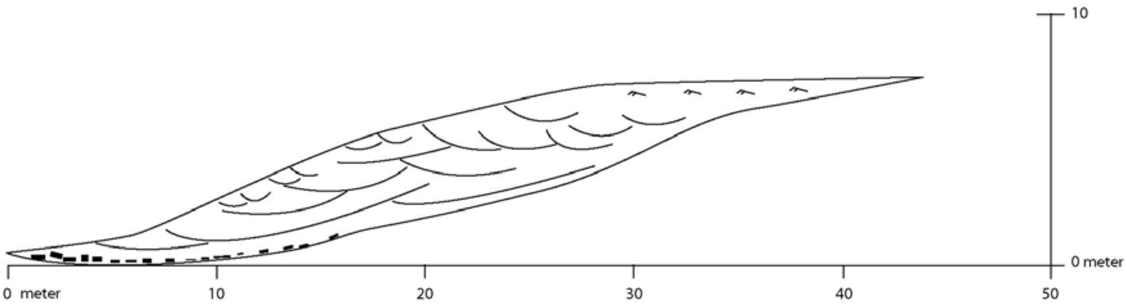


Figure 6.3: Facies development of one single well preserved clinotheme, lateral accretionary element from the lowermost channel elements in panel 18.

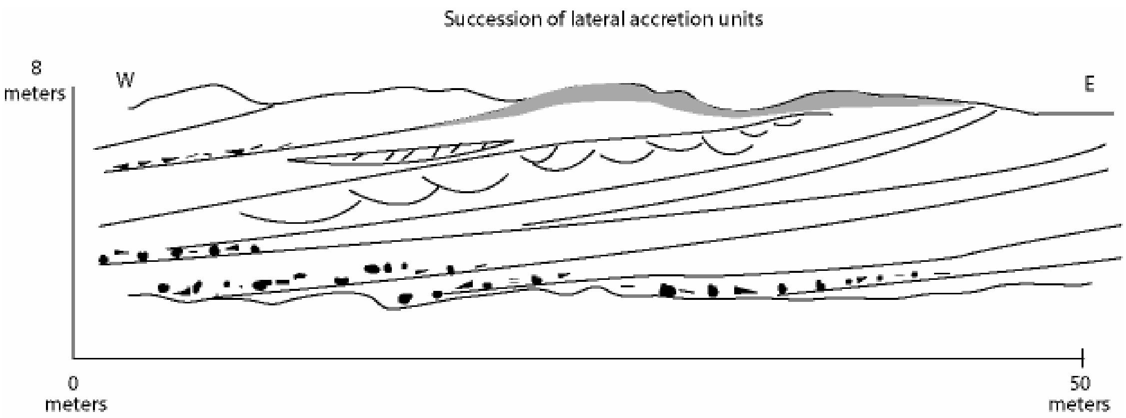


Figure 6.4: Succession of LA units from log 17.2.2.

Interpretation

Lateral accretionary bars are deposited as a result of helicoidal flow patterns set up where the main flow of the channel is directed away from the bank (Miall, 1996). It is most common in sinuous channel reaches on the inside of the channel curves, in meandering channels LA units are termed point bars. Lateral accretion units may also be deposited in braided or almost straight channel reaches with slight internal meandering as secondary helicoidal flow is set up by bank-attached or mid-channel bars (Collinson, 1996).

Helicoidal flow patterns developed in sinuous channels (figure 6.5) leads to erosion of the outer thalweg while lower flow energy and shear strength leads to deposition on the inner bank. The flow lead to a general upward fining pattern from lags on the lower clinothemes to fine sand and silt on the top, and facies development in a lateral accretion unit reflects an upward declining flow power and shallower depositional setting (Allen, 1984). Mud fragments and calcrete pebbles in conglomerate deposited on the lower part of the clinothemes (toesets) are swept up from the channel floor by the helicoidal flow. The lithology of the clasts in the Colton Formation reflects local conditions, as calcrete and mudstone clasts from paleosol and mud chip lags derived from dried-out parts of the channel surface itself. Calcrete clasts are rarely brought all the way to the top of the clinothemes. Mud chip lags are occasionally found deposited all the way to the top, indicating that the clasts may have experienced virtually no transport after mud drapes were deposited on the channel and clinotheme during falling and low stage.

Clinothemes build out at an angle to channel flow direction by dunes migrating in the direction of flow across the lateral bar surface (Allen, 1984). The angle of the lateral accretion to the flow direction of the channel is varying due the combined effect of both lateral and downstream erosion of the channel banks (Miall, 1996). If the downstream translation of the channel curves or bars is small, lateral accretionary elements may be building out perpendicular to channel flow, even with some accretion opposite to the channel flow. The relative importance of lateral and downstream migration of lateral bars may be reflected in the grading of vertical successions. Bridge and Tye (2000) stated that although most channel bar and channel fill sequences are upward fining, dominant lateral expansion leads to better preservation of upstream parts of bars. Upstream parts of bars tend to have little vertical variation in grain size and downstream bars tend to be upwards fining. Dominant downstream translation and thus preservation of the downstream part of bars leads to upward fining successions. If their principle is applied to the Colton outcrop, lateral movement is of larger importance in the lower part of the central channel belt than in the upper part. Grain size development in vertical successions can indicate the angle of migration to paleoflow.

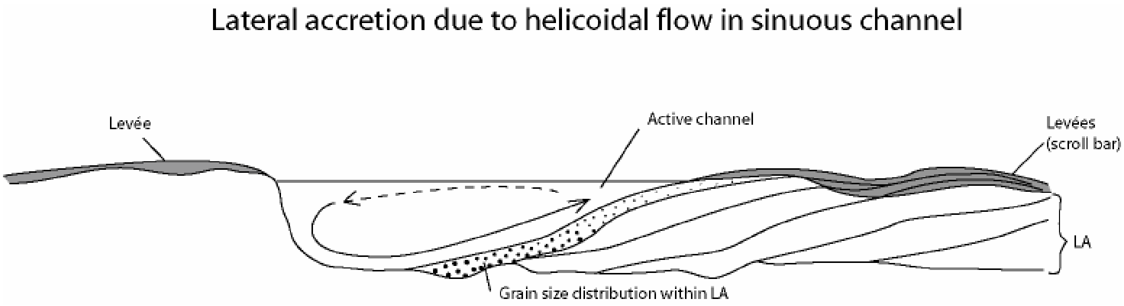


Figure 6.5: Helicoidal flow, LA and overbank.

The height of LA units is identical to channel depth and LA unit length is dependent on channel width. Angle of dip of the LA unit surface reflects the steepness of the channel and thickness of LA unit reflects the amount of accretion per flood episode, and hence the lateral migration of the channel (Bridge and Tye, 2000). The channel from panel 18 in Fig 6.3 by this principle shows larger lateral migration per flood episode than the LA units in log 17-2-2 (figure 6.4). The reason for this may be larger water discharge and thus larger erosion.

6.2.3 Downstream accretionary bars

Description

Not many elements have clearly been identified as downstream accretionary bars. Towards the top of log 18-2 (361 – 368 meters), a good example is seen (Fig 6.6). Lateral extent of bar is 5 meters in outcrop, but the whole bar length is not exposed. It consists of 2.3 meters of tabular cross stratified sandstone facies (B1) overlain by 2.2 meters of trough cross stratified sandstone facies (B3) in bed of 0.2-0.5 meters thickness. The B3 beds have downstream dipping bounding erosional surfaces. Dip of cross-bedding and bounding surfaces are approximately in the same direction.

Interpretation

The facies of the downstream accretionary bars indicate that there were both 2D and 3D-dunes migrating downstream. Downstream migrating bars are elongated parallel to the direction of flow and not connected to the channel sides. They are found in all channel types.

Downstream accretion bars height reflects channel depth (Miall, 1996). There is a contrast of thickness between the facies, which may reflect that the underlying 2D-dune is a stranded bar with small 3D-dunes superimposed as a response to low water stage or vertical build-up of bar. The reactivation surfaces indicate erosion by rapid falling water level, as a general principle suggested by Collinson, (1970).

6.2.4 Braided bars/sandflat

Description

The channel bodies 4, 11 and 19 in panel 17 and 18 (Fig 3.3) features what is interpreted to be sandflats. Sandflat facies association are composed of low angle trough cross stratified (B2), trough cross stratified (B3) and cross laminated sandstone facies (B6). The facies association is characterized by interfingering low relief bars (2-4 meters height) with a limited lateral extent of single bars but a total length in outcrop of up to 500 meters (channel body 19). Bounding surfaces within the sandflat is bedform, interbed or reactivation surfaces, and mud chip conglomerate is found above erosive bounding surface. Bars often have a distinct upwards fining pattern within each bar and may contain plant debris (log 18-3-3, 152 meters).

Interpretation

Sandflats are reaches within a low sinuosity or braided channel of overlapping and migrating bars which may be exposed and eroded at low stand. They are bars shifting rapidly over a shallow reach of the channel. Sandflat bars may accrete dunes upstream, downstream and laterally (Collinson, 1996). Plant debris might represent vegetation on the bars, or drifting plant material being trapped on the sandflat when they are partially exposed.

As mentioned above, upwards fining pattern is more likely to be found in the downstream part of the bars, and this strengthen the assumption of channel 11 in log 18-2 and channel 4 in 17-2 as sandflats consisting of bars migrating downstream.

6.2.5 Abandoned channel infill

Description

Infill of preserved channel topography if found in at least four places in the panels. Abandoned channel infill bodies are closely associated with LA units and have a sharp and steep erosional channel floor surface cut into surrounding facies (Fig 6.7). Measurements of these bodies are shown in table 6.3 in section 6.2.6. The infill sediment of the channel space is rather homogenous and sand dominated, consisting of calcrete nodules/mud clast conglomerate (A2) towards the bottom, small scale low angle trough cross stratified (B2) and trough cross stratified (B3) sandstone, but mostly structureless sandstone (B7), and sandstone facies' are fining into floodplain fines in the very uppermost 1-2 meters. 1-2 meters of floodplain fines at the top might however be an underestimated number, since the upper boundary of the thalweg is not always distinguishable. Scours are found at the lower boundaries of abandoned channel infill.

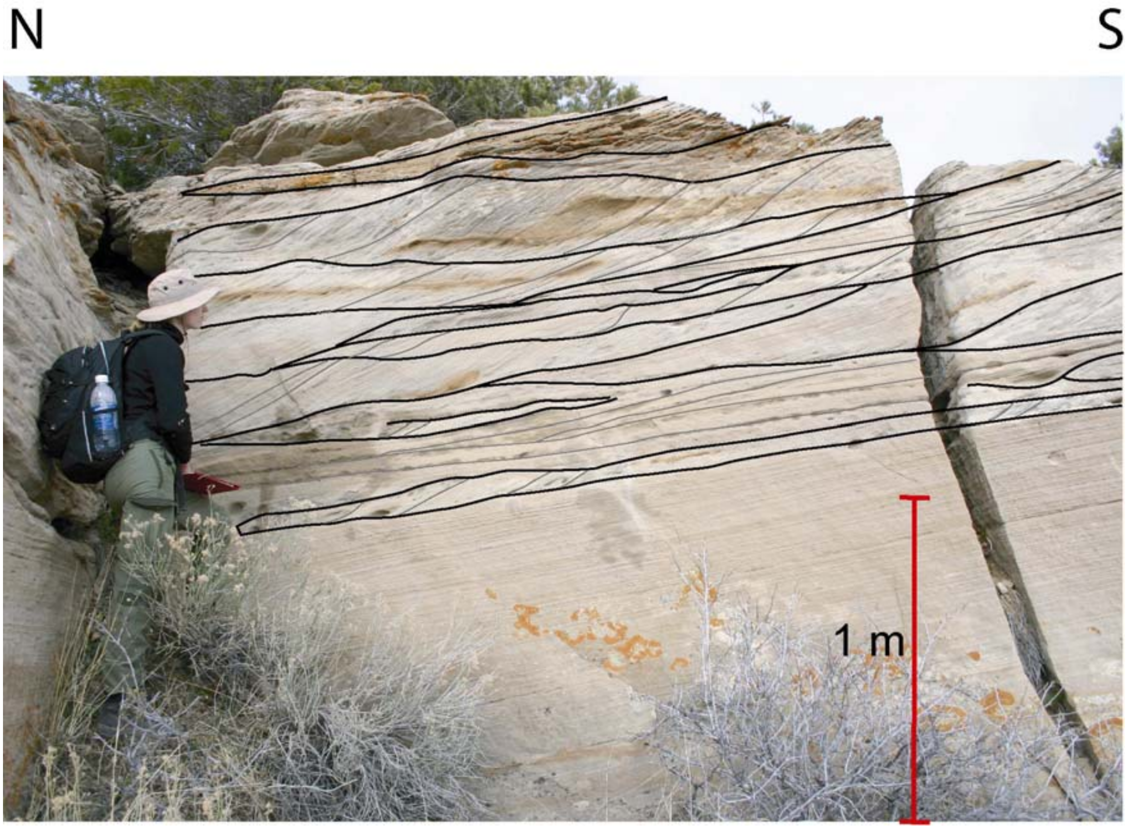


Figure 6.6: Downstream accretionary bar in log 18-2 (361 meters). Migrational direction of dunes to the north. Note downstream dipping reactivation surfaces on the uppermost small dunes.

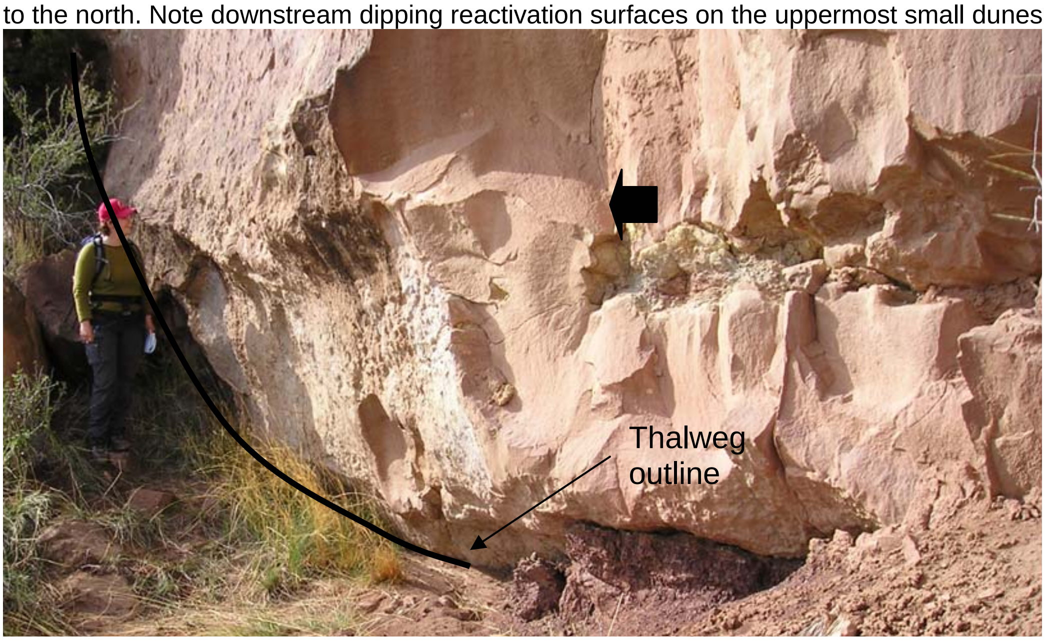


Figure 6.7: Channel 5 cutting into floodplain fines, panel 17. Channel flow direction was 20° into profile, as indicated by arrow. Note sandy abandoned channel infill.

Abandoned channel infill with concave upwards both upper and lower boundaries is found in the panels as channels 20 and 32, and a segment of channel 3. Channel nr 20 is 130 meters wide and 15 meters deep, and channel nr 32 are 150 meters wide and 20 meters deep. In channel 3, the infilled incision is 50 meters wide and 6 meters thick. Erosive boundaries within the units are reactivation surfaces.

Interpretation

This type of infill of channel space is a result of channel abandonment. Channel abandonment is caused by avulsion or neck cut-off by meander bends. Channel avulsion is a consequence of channel belt building alluvial ridges above the floodplain because overbank sedimentation of fines creates cohesive channel sides. As alluvial ridge build up higher, the situation is increasingly unstable relative to the lower-lying floodplain. Channel banks are eventually breached by flood and active channel segments are relocated to the lowest lying part of the floodplain.

According to Bridge and Tye (2000), sand rich abandoned channel infill generally is an indicator of low sinuosity channels. Low angle between abandoned and new channel leads to continued flow in the older channel after establishment of a new flow, particularly in flood periods. The old channel receives ephemeral sand input for a period until the upper part of the channel is clogged, and the channel path is truly abandoned and filled in by fines.

The convex-upwards channels are most likely deposited in pulses of reduced flow strength. A large initial erosional incision is infilled by lateral accretion units and channel-floor bars from channels of successive decreasing depth, generating the reactivation surface which follows the channels inner boundary. The incised channel infill in channel 3 is most likely from a chute channel, since it is shallower than the channel body it cuts into (Bridge and Tye, 2000).

6.2.6 Channel infill

Description

Singular channel infill bodies can be followed from 60 meters up to 650 meters within the panels, the average value is 220 meters in lateral extent in the studied outcrop. Height is ranging from 4 to 15 meters, but is commonly about 8 meters and rarely exceeds 10 meters. Lower boundary of channel infill is sharp and erosional above all other facies' and facies associations, sometimes with deep scours. The lower boundaries of singular channels infill elements show up to 20 meters large scale erosive relief cut into underlying sediments. Upper boundary is a conform surface below floodplain fines or levees facies associations, but sometimes appear transitional. Within the channel belt, the boundaries are erosional channel floor surfaces (of table 6.2) between channels. An example of a channel infill can be seen in figure 6.8.

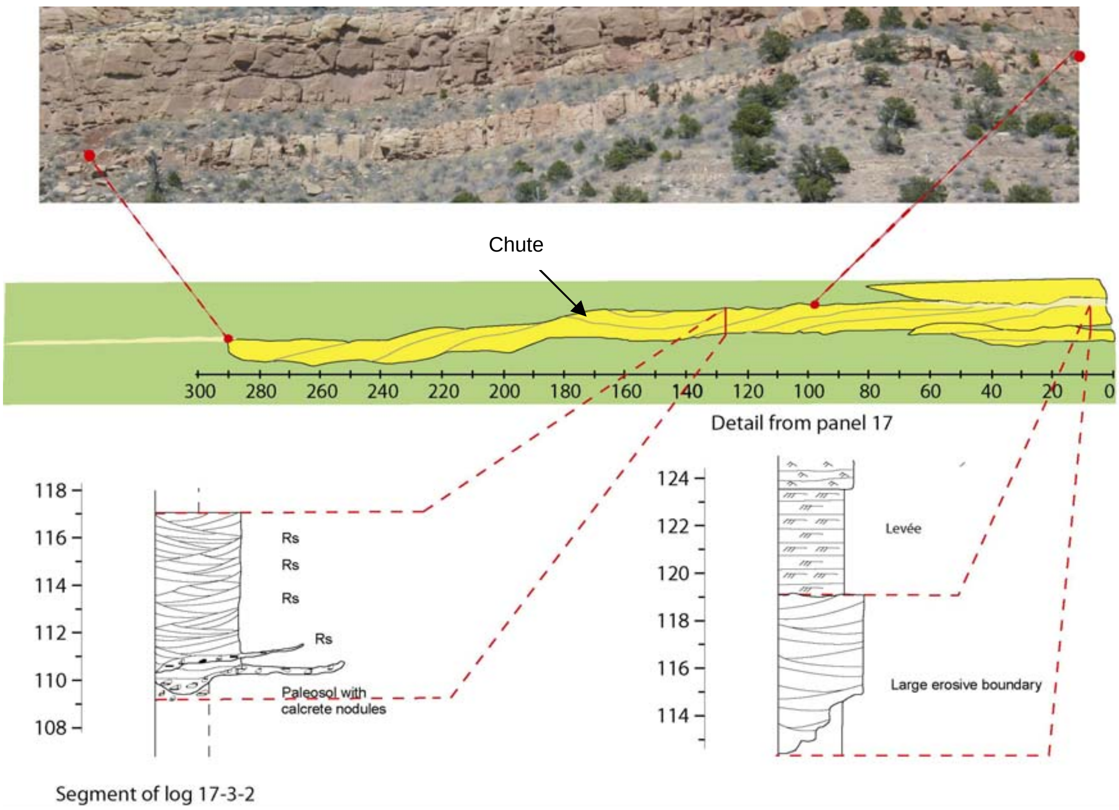


Figure 6.8: Single laterally migrating channel infill from panel 17. Note chute channel infill to the left of log 17-3.

The internal geometry of channel infill, when identified, mostly consists of laterally stacked lateral accretionary elements. Towards the bottom of channel infill elements lenticular bars can be found, both consisting of a single facies, as calcrete nodule/mud clast conglomerate (A2) or trough cross stratified sandstone (B3). Lateral accretionary elements build out without being cut off by erosion for substantial length of channel migration, even in the central channel belt.

External geometry of the channel infill sandstone bodies is concave upwards (ribbon sandstone), laterally restricted sandstones with lateral accretion and multistory and multilateral. In the lowermost and uppermost part of the outcrop, mostly multilateral single storey or single channel bodies can be found, but in the lower and upper central parts of the outcrop the channel bodies are multistory and multilateral. The amalgamated zone in the lower central parts is termed the central channel belt. It consists of approximately 16 multistory and multilateral channel infill bodies. The thickest vertical succession is found within panel 18, where up to 7 vertically stacked channel bodies forms a sandstone cliff over 50 meters high. Internally within the central channel belt, there are found mud drapes between channel bodies.

Of the 38 identified channel belt sandstones, 4 were found migrating broadly east, 12 were found migrating broadly west and the rest was found to have elements migrating in several directions or had no clear migrational direction in the panels. Broadly east and broadly west should be interpreted as north and south respectively, when the channel paleoflow direction in the profile is taken into consideration.

The nature of the conglomerates within the channel sandstone bodies varies laterally and vertically. In panel 17, only the lowest 10 meters of channel infill has calcrete nodules in their conglomerate, but in panel 18 there are calcrete nodules in the conglomerate in the middle of log 18-1 and 18-2. Erosion has at some places been so large that the upper, fine grained or clay-drapes parts of the channel infill have been removed. This does not occur at the same height in the channel belt in all the logs. Preserved levées are found above channel infill in the upper part of logs in panel 17, but in panel 18, levées are more preserved in the lower part of the logs, but exist throughout.

Channel abandonment infills are generally having a small width/depth ratio, stemming from deep, narrow channels. Trunk dimensions can also be estimated form lateral accretionary elements. A table of estimated channel trunk dimensions can be seen below in table 6.2. W/h ratios of channel sandstone bodies are discussed in chapter 8.

Table 6.2: Estimates of channel size based on presumed uneroded channel infill sandstone bodies from panels. Based on assumptions from Miall (1996) of that height is identical to bankfull depth and width is 2/3 of lateral extent of lateral accretion units. No height correction made for compaction. When flow direction is unknown, general paleocurrent direction of 305° for Colton is applied, see chapter 7.

Abandoned channel infill depth	**Abandoned channel infill width**	**Flow direction**	**Orientation of profile**	**Channel infill height**	**Estimated channel width**	**Estimated w/h**
9 m	23 m		300°	9 m	22 m	2,4
7 m	35 m		300°	7 m	33 m	4.7
9 m	50 m	330°	350°	9 m	46 m	5,5
9 m	?			9 m		
LA height	**LA width**					
15 m	60 m	330°	240°	15 m	60 m	4,0
5 m	20 m	330°	350°	5 m	18 m	3,6
20* m	70 m		300°	20* m	67 m	3,4
* Minimum height, as channel might be eroded by overlying channel bodies						

Interpretation

Channel infill facies association is dominated by lateral infill architectural elements. Architectural elements were deposited as bars in different parts of the channels as the channels migrated over the alluvial plain. In a laterally migrating channel, the lateral accretionary bars connected to the channel side opposite of the migration direction have the best preservation potential. Downstream accretionary bars can also be incorporated into the migrating lateral bars and preserved.

According to Harms (1982) sandstones tend to be connected laterally and vertically when avulsion rate is high, regional subsidence is low and streams are of low sinuosity. Intervals dominated by single or single storey channel sandstone bodies with limited lateral extent may reflect channels with stable banks, which can be both meandering and low-sinuous. Such intervals must however also be investigated with respect to accommodation space generation and floodplain stability, which will be further discussed in chapter 7.

Incision by active channels seems is seen to be on several scales. Undulating lower boundaries of single channel infill sandbodies can have over 10 meters erosive relief as seen for instance in channels 3, 18 and 24. According to Dalrymple et al. (1994), such undulations should be considered as normal results of channel avulsion, stream capture and base level fall. The lower boundary of the lower channel belt, however, has an erosive relief of 20 meters. Laterally it extends longer than the infill of single channel infill sandbodies and incision is deeper than the channel infill height, so it is possible that it could be interpreted as an incised fluvial valley (Dalrymple et al. 1994). The cause of such valley incisions may be both base level fall, uplift of source area and increasing discharge (Dalrymple et al., 1994).

When the rivers have eroded into floodplain fines, there are often calcrete clasts, but within the central channel belt, many rivers were eroding into a mainly sandy alluvium consisting of previous channel sandbodies. The presence of calcrete nodules in conglomerate and the levées indicate that there have been variations in channel activity laterally and vertically. In addition to these variations, lateral migration seems to have a preferred westerly migration. The causes of these variations may be climatic, tectonically influenced, or caused by paleotopography or subsidence, and will be further discussed in chapter 7. The estimated w/h ratios from the table 6.3 are very low and indicate that the banks of the channels were stable, even within the central channel belt.

6.3 Facies association II: Overbank

The facies association is subdivided in levée and crevasse channel. They are usually found adjacent or closely associated with channel bodies. Overbank sediments were deposited as rivers during high stand splashed sediments up on and over the channel sides. The rivers that were the sediment sources are not always exposed in the profile.

6.3.1 Levée

Description

Levées consists of small structured through cross stratified sandstone (B3), planar stratified sandstone (B4), planar laminated sandstone (B5), cross laminated sandstone (B6), structureless sandstone (B7), cross laminated siltstone (C1), planar laminated siltstone (C2), structureless siltstone (C3) and claystone (D) facies'. The most common facies are massive or rippled silt/fine sand with high clay content that are deposited as thin drapes or in bedsets ranging from 0.3 to 6 meter thickness. The individual beds in the bedsets are 0.2 - 0.5 meter thick, but beds are often indistinguishable due to bioturbation. Lateral extent of single levées is 20-30 meters, in successions of overlapping levées this facies association may extend for 150 meters. The lower boundary is often transitional to facies in lateral accretion units, or found draping the units, and in outcrop may appear as a continuity of the lateral accretion units (figure 6.9). In some cases, levées have mud chip lag at the lower boundary. Laterally, levées pass into floodplain fines, and bedsets of this association may form wedges thinning and fining away from channel bodies.

Levées are identified in outcrop usually on the base of the geometry of the element, the red color and the association with channel and channel infill elements vertically and laterally. Levées can be fine grained; silt or clay, but it is not useful to base interpretation of this facies association on grain size or structures. Often the levées in the Colton Formation consist of sandstone, very fine or fine, and structures have been destroyed by bioturbation. Even the sandstone levées do, however, have high clay content and red coloring.

Levées within the central channel belt had low preservation potential, and were often eroded by overlying active channels. The result is that the levee facies association is highly discontinuous in outcrops. The best preserved levee facies association elements are found to the top of the channel belts, and are up to 6 meters thick.

Interpretation

Levées are deposited on the channel banks during flooding by materials in suspension in the river. The material is usually fine grained because the finest materials are found in suspension in the upper part of the river, but strong floods may bring coarser material. Levées are red because they are not submerged except during flooding and are exposed to bioturbation.

6.3.2 Crevasse channel

Description

Crevasse channel infills are represented by 1-4 meters thick deposits, usually deposited in close proximity of main channel bodies. The channel infill successions have concave upward erosive lower boundaries when viewed perpendicular to flow direction. Some crevasse channel infills have rip-up clasts towards their lower boundary. Facies in the channels usually show relatively high energy, as low angle trough cross stratified (B2) and trough cross stratified (B3)

Figure 6.9: From bottom; clinotheme which is upwards fining at the top with an overlying red-colored levee. Above levée an erosive boundary below new channel element. The levee in the picture has an lateral extent of 8 meters before it is truncated by channel sandstone body in the far right in the picture.

Figure 6.13: Erosive boundary between underlying channel infill and overlying beach sand, from log 18-2-3 (133 meters).

sandstone, deposited in 0.2-0.3 meter high beds. In 18-3 the thickness of the crevasse channel is 2 meters and exposed width is 15 meters. There are not found many crevasse channels within the Colton Formation, but a crevasse channel infill sandstone of thickness 4 meters is found in panel 18-1-1.

Interpretation

Crevasse channels incise into channel bank or levées due to water overflowing and eroding the banks during flood. A 4 meter deep channel indicates that there was a gradient between the bankfull river and the surrounding floodplain, allowing the crevasse channel to incise deeply into the bank. This crevasse channel could well be reactivated by several floods, and such deep crevasses may ultimately lead to channel abandonment by developing into the main channel.

A relatively steep slope gradient from the levée of a river down to the floodplain usually indicate that there are a lot of mud in suspension in the river, causing the formation stable banks and the establishment of the channel belt above the floodplain.

6.4 Facies association III: Floodplain fines

The floodplain fines are divided into three sub-facies associations; mudstone, crevasse and paleosol. Floodplain fines form the most substantial part of the Colton Formation, and within this facies association, mudstone is dominating. Thick vertical and lateral successions of mudstone are found within the formation, and within the mudstone there are irregularly spaced sheets or lenses of crevasse splays.

Sediments in floodplain deposits are provided by floods overtopping river banks or by local sheets floods or slumps on the alluvial plain, in some cases due to heavy rainfall. They are usually fluvial, but in a dry climate with sparse vegetation, some sediment may be wind-blown. Unless aeolic sediments form dunes, they are perhaps indistinguishable from fluvial deposits, the poor quality of the exposures taken into consideration. Sediments are not provided continuously, floodplains have a sporadic sediment supply in between long inactive periods. The lateral and vertical variations within the floodplain fines are strongly affected by channel proximity, channel avulsion rates (e.g. Bridge and Mackey, 1993) and incision. It affects paleosol maturity and successions, sandstone percentage within the fines and density of individual crevasses in a vertical succession.

6.4.1 Mudstone

Description

Mudstone is used as a common term for structureless (C3) and claystone (D) facies, of which C3 is the most common. Within the panel, mudstone shows a vertical continuity for over 100 meters and a lateral continuity throughout the panels. Laterally and vertically there are transitions between siltstone and claystone. Upper boundary is erosive; deeply erosive channel floor or channel belt floor below channel bodies, less erosive below crevasses. Mudstones are commonly red, but also grey (especially facies D).

Interpretation

Mudstone represents vertical aggradation on the floodplain of distal splay deposits, sheet floods or as floodplain pond sand. The sediments reflect a low energy setting. Color of the mudstone may reveal lateral and vertical variations in depositional settings and sediment supply, as discussed in paleosol facies description (chapter 5). Coloring may also be connected to clay content, as permeable, sand-rich sediments are colored by precitipation of hematite quicker than clay-rich sediments (Collinson, 1996).

The floodplain fines were probably rather cohesive even relatively shortly after deposition due to the clay content, as is indicated by steep cutbanks.

6.4.2 Crevasse splay

Description

Crevasse splay is exposed as vertically stacked or single sandstone sheet units and lenses consisting of low angle trough cross stratified (B2), through cross stratified sandstone (B3), planar stratified sandstone (B4), planar laminated sandstone (B5), cross laminated sandstone (B6), structureless sandstone (B7) facies' of which B7 is the most common facies. Lateral extent may be large, varying from 4 to almost 400 meters. The sheets consist of single beds or bedsets with small internal erosive or conformable bounding surfaces, often with a lateral transition of facies within and between beds. Bed surfaces may be undulating, and beds are also found as lenses within bedsets. Bed thickness is usually 0.2-0.3 meter and maximum sheet thickness is 4 meters. Red coloring and bioturbation is common, but they may also be yellow or grey.

The sheets can laterally be a transition from levées or channel bodies and ultimately show a transition to floodplain fines. Quite often there is no clear transition from channel sediments exposed within the profile. Lower boundary can be erosive (also with scours) or conformable, and upper boundary can be eroded beneath all overlying sandstone facies or be conformable below all fines. Within the fines, the upper boundary may be undulating.

Bed and bedset frequency and thickness is largest in the vertical succession in the proximity of channel bodies and gradually becomes thinner and usually only isolated beds within the floodplain fines the further away they appear from channel packages.

Interpretation

Crevasse splays are sediments fed from crevasse channels; deposition is caused by loss of competence because of flow expansion and loss of flow power.

Bed thickness may reflect size of flood, amount of sediments in suspension and proximity to channel. Vertical frequency reflects channel proximity, since it is clearly seen that the frequency is highest above, below and laterally adjacent to channel bodies. An increase of number of beds and bed thicknesses may be regarded as 'forebodings' of an approaching migrating channel, alternatively an avulsion. Undulations and beds overlapping laterally may be the result of crevasse splay progradation caused by migrating channels. Isolated lenses are typically smaller and may be distal or small crevasse splays. Undulating upper surface probably reflects preserved dune topography within the bed.

6.4.3 Paleosol

Description

Paleosol occurs in the mudstone facies association. Lateral extent of paleosol is up to 400 meters, but deep channel scours often truncate their lateral extent. The partly massive calcrete layer, seen in log 17-2 (15 meters) has an upper boundary below channel infill sediments, and seems to have prevented the channel from eroding further down, since base of channel follows the upper boundary of the calcrete layer. Some of the paleosols seems to be of a large lateral extent, but others are more local. Relief of paleosol is maximum 15-20 meters.

Interpretation

Paleosol with calcrete development reflects a paleoland surface which has been stable for a significant period of time with a very limited sedimentary input. It may be linked to periods of large fluvial incisions where there has been paleosol development on terranes. Development of the calcrete below the central channel belt can support a theory of large incision prior to deposition of the central channel belt sandstone infills. Profile of paleosol indicates either the paleotopgraphy, in which case erosion and perhaps incision has taken place before the paleosol development, or later differentiated compaction of the alluvium.

Paleosol horizon development is a function of both time available for maturation, space available for channel migration and available sediments. Lateral variations in paleosol pedofacies are expected to be increasingly maturity of paleosols with increasing distance from the active channel (Retallack, 1988). Immature paleosols are expected to be found immediately adjacent to the channels. Thick sections of red colored silt with calcrete nodules are interpreted to be cumulative soils (Kraus and Brown, 1988), and the red floodplain mudstone can be regarded as cumulative immature paleopeds. Cumulative soil development reflects a slow but steady sediment input (e. g. Kraus and Brown, 1988).

6.5 Facies association IV: Lake

Lake sediments are subdivided into lacustrine mudstone and beach facies. All the three facies associations have a large lateral extent, but not necessarily throughout both the panels. It is not straightforward to identify lake sediments, as the facies found are similar to the Colton Formation sediments. Within the sandstone facies, lacustrine facies is interpreted primarily based on the presence of *Scolithos* trace fossils in the *Scoyenia* ichtnofacies assemblage.

Lacustrine mudstone or sandstone found interfingering with fluvial sandstone of the Colton Formation is most likely Green River Formation sediments deposited at highstand levels of the Lake Uinta sediments, but they may also be deposited in floodplain ponds. Floodplain pond development may also be connected to the lake water level, since a high water level in the lake will heighten the ground water level and enhance floodplain pond development.

Total vertical span of occurrence of sediments I have interpreted to be lacustrine can be seen in table 6.3. In all the logs, except 18-1, there is a zone ranging from 28-50 meters below datum that is interpreted to contain lacustrine sediments along with the continental floodplain sediments. In 17-2 and 18-2 there is also evidence of a lower, lacustrine influenced zone. This differentiation of heights of lacustrine influence is however very constructed. There is no reason to believe that there has been a paleotopography during deposition of 70, but perhaps 20-30 meters on the alluvial plain. Nevertheless, lacustrine influence can be assumed for the whole section above the lowermost occurrence observed.

Table 6.3: Extent of possible lacustrine facies below top datum.

Log	**17-2**	**17-3**	**18-1**	**18-2**	**18-3**
Beach (m)	6	5,5	0		3,5
Crevasse splay fan (m)		7	0	5	1
Mud, green or grey (Y/N)			Y	Y	
Height of lacustrine occurrence (m)	28 (101)	35	?	50 (88)	36
Number of non-lacustrine crevasses in zone	1	0		0	1
Comments	Zone 101-93 m below datum with green/grey mudstone with some red layers. Datum erosive above channel infill.	Soft sediment deformation, water escape structures and rip-up clasts in crevasses		Datum with erosive lower boundary above channel fill. Crevasse 1 meter above Datum.	

6.5.1 Lacustrine mudstone

Description

Mudstone thought to be of lacustrine origin is found below the datum in the outcrop. Thin, green clay dominated mudstone laminations in 2-10 cm thick beds within red floodplain fines may be present in up to two meters thick bedsets, but were not confidently established in the field. They are found within an extensive package of fines.

Interpretation

Lacustrine mudstone is deposited from suspension in lacustrine environments of low hydraulic regime.

6.5.3 Beach

Description

The beach deposits consist of low angle trough cross stratified sandstone (B2), trough cross stratified sandstone (B3), planar stratified sandstone (B4) and planar laminated sandstone facies (B5). They are set apart from lacustrine crevasse splay fans by the high-energy facies and the low angles of cross stratification. They are laterally extensive sheet-like bodies of sandstone, up to several hundred meters long and less than 4 meters thick. Laterally beach deposits have an interbed boundary to mudstone. Channel deposits are found below lacustrine deposits bounded by sharp, mostly planar erosive surfaces as seen in figure 6.13 and log 17-2-3 (101 meters) and log 18-2-3 (133 meters).

Interpretation

Beach deposits are limited in vertical extent. The vertical thickness is limited by the lake level stability and amount of sediment available for reworking in the beach zone. Since the gradient of the system is so low, small changes would rapidly lead to changes in the depositional setting, thus the beach zones might not have been stable for a long period.

Beach deposits are formed by the swash and backwash of waves in the beach zone, reworking underlying sediments, or sediments supplied by coast-parallel transport or deltas.

7 Depositional environment

7.1: Depositional setting of alluvial plain and lacustrine sediments

Colton Formation sedimentary rocks represent a range of fluvial sediments from depositional environments extending from medial alluvial plain to fluvial mudflats on the distal alluvial plain (figure 7.1 and 7.2), as well as lacustrine mud. Medial alluvial plain sedimentary environment consisted of channels, crevasses and floodplain mudflats. Distal alluvial plain consisted of fluvial lobes of shallow distributary channels that were terminal with respect to bedload coarse clastic sediments, and floodplain mudflats which towards the shoreline were transitional to lacustrine mudflats. The lack of coarse clastic delta sediments in the sedimentary record indicates a low competence of the fluvial system, which can be expected due to the low slope angle of the southern lake margin of Lake Uinta. At the type locality for the formation in the Colton area located 30-40 kilometers further to the west there are several good outcrops of isolated distributary channel sandstone bodies, (Fig. 7.3).

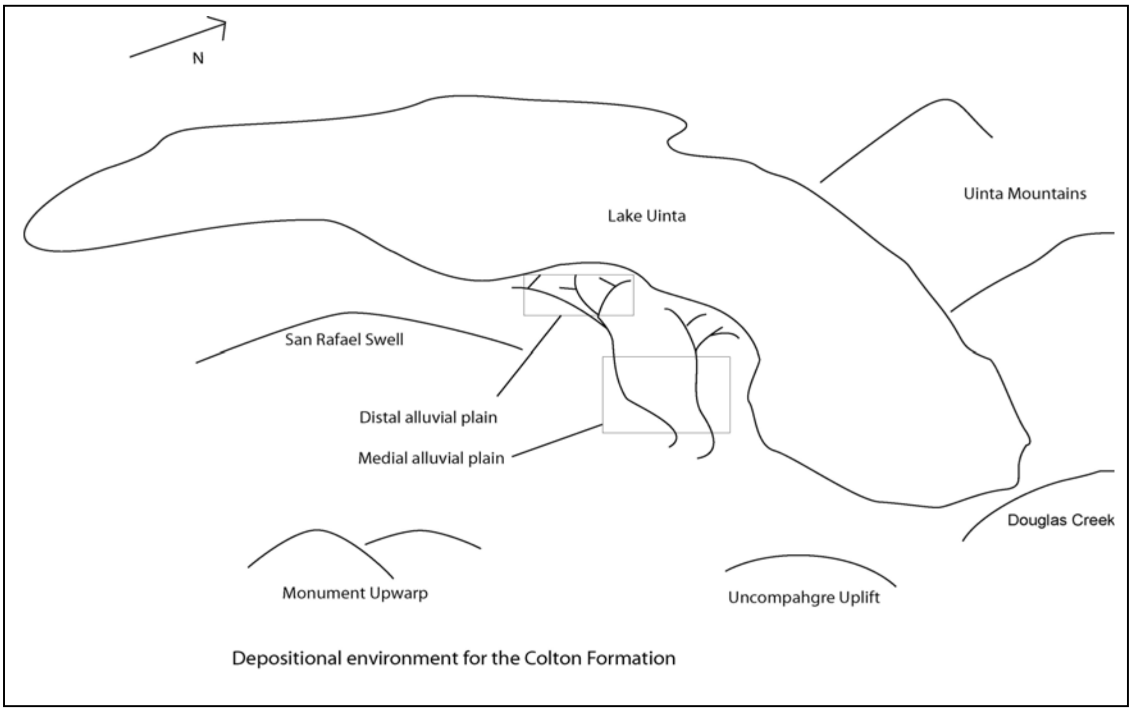


Figure 7.1: Position of the Colton Formation in Lake Uinta.

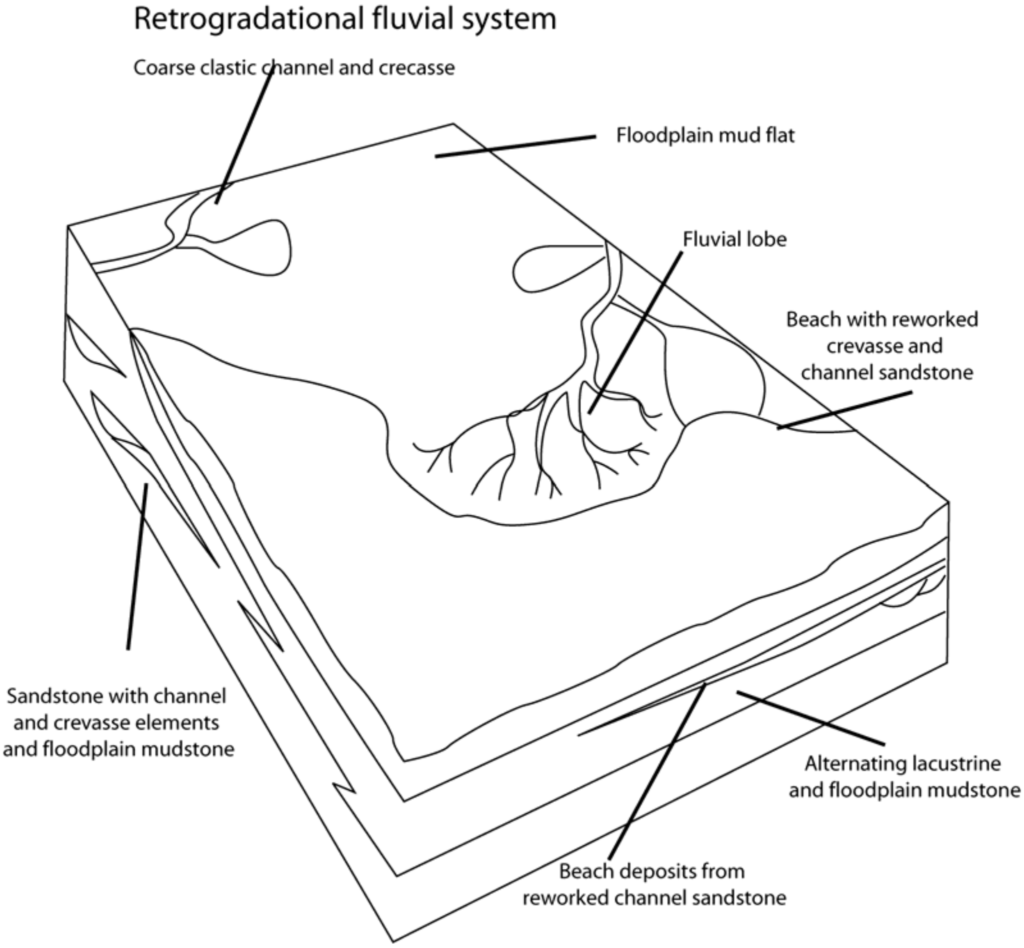


Figure 7.2: Depositional model for the Colton Formation in a retrogradational setting.

The fluvial system of the Colton Formation must have created a very slightly sloping alluvial surface, at least in its lower reaches along the ramp type southern basin margin of the Lake Uinta. The mudflats in the fluvial/lacustrine contact zone contributed to preserve a low slope gradient out into the lake since mud has a very low critical angle. The competence of the fluvial system was also dependent on the paleoslope, but probably to a much larger degree by the discharge. Discharge reflected size of and climate in the drainage area. In the early basin stages, the drainage area may have been small and segmented yielding less sediment than during the later, more mature stages of erosive relief.

No extrabasinal clasts above medium sand size found in the sediments. Explanations to this may be the previously mentioned low competence of the river system or that the sediments are sourced by erosion of older sedimentary rocks (Stanley and Collinson, 1979) that did not give rise to debris larger than the dominating medium sand fraction.

Several factors indicate a confinement of the area that was available for channel migration. The general paleodrainage direction from south-east to north-west differs from the basin axis, which is broadly east-west. Space available for development of a traditional large fan delta must have been very restricted. Mulitstorey and multilateral channel sandbodies implies that the available space for channel migration was restricted. Possible explanations to this confinement could be valley incision or topographic rises. As previously mentioned in chapter 6 the central channel belt shows indications of being an incised valley infill. Total extent of the central channel belt sandstone body estimated by the SAFARI project is 4 kilometers lateral to paleocurrent flow and some tens of kilometers parallel to paleodrainage direction (SAFARI, 1995). It is not an unlikely extent of a fluvial incised valley, as described by Schumm and Etheridge (1994). Another explanation for the confinement may be that the depositional system was bounded to the east and north-east by the Uncompahgre Highlands.

Assymetric tectonic subsidence events probably affected the river courses, either by forcing the streams to migrate laterally into direction of maximum subsidence, or/and by imposing a new pattern of avulsion. As noted in the facies association chapter, many lateral migrating channel bodies shows a southwestern migrational direction in the outcrop. According to Alexander and Leeder (1987), tectonically affected lateral diverted flow will lead to preservation of lateral migration elements and abandoned channel bodies which show a dip up the paleoslope. If the northwesterly oriented river system was affected by tectonic subsidence in the north, the observed southwesterly preserved lateral migration units in the Colton Formation should be expected.

Figure 7.3: Distributary channel sandstone at Kyune, in the type area of the Colton Formation.

7.2: External controlling factors of base level

The concept of the base level denotes an undulating equilibrium surface dividing erosion and deposition (e.g. Shanley and McCabe 1994). The stream equilibrium profile (Dalrymple et al. 1998) makes sequence stratigraphy more easily applicable to fluvial systems. The stream equilibrium profile is the hypothetical stream profile that is thought to instantaneously adjust to the interaction of all controlling factors. The balance between erosion, sediment bypass and net sedimentation is controlled by the stream equilibrium profile (graded stream profile). The equilibrium surface is termed the *pseudo base level* within this concept (Dalrymple et al. 1998).

Important controlling factors on the base level in are tectonic subsidence and uplift, lake level, discharge and sediment supply, in accordance with the factors listed by Shanley and McCabe (1994). These factors interact, and have varying degrees of impact in the different parts of the depositional system.

Lake level

The Uinta Lake has acted as the terminal base level for the Colton Formation fluvial system. Impact of variations in lake level is generally that a fall in base level leads to erosion, and a rise leads to deposition. The impact of lake level fluctuations will be large on the distal alluvial plain and diminish gradually further up in the alluvial system. In the alluvial shelfal areas of the large Lake Uinta, the stream equilibrium profile has been strongly affected by lake level changes.

Tectonics

Tectonic subsidence in the basin leads to a lowering of the lake level. Uplift of the hinterland increases the paleoslope and the profile of the catchment area. Hinterland uplift may also affect the climate, creating a wetter hinterland climate leading to increased discharge. Hinterland tectonic uplift by this leads to larger volumes of sediments to be fluxed into the system. Increased discharge and sediment supply enhances the effect of the increased paleoslope and leads to base level being moved further down the paleoslope.

Lake level fluctuations were induced by climatic variations and tectonic subsidence of the basin. Lake level could have fluctuated quickly and dramatically by these reasons. There would probably be a time lag in the fluvial system's response to changes in base level and paleoslope, and the short fluctuations of the lacustrine base level will only be recorded in the lowermost part of the alluvial system.

Climate

Climatic changes are often linked to the Milankovitch cycles of 11 000, 22 000, 41 000, 100 000 and 400 000 years. Such high-frequency climatic cycles are generally suggested to exert marked impacts on lake levels, particularly in possible semi-arid regions like that of the Eocene Uinta Lake (Keighley et al., 2003).

Tectonically induced increase of paleoslope along with an associated fall of lake level as the lake are deepened would increase sediments transport and lead to a progradation of the Colton Formation out into the Uinta Basin. As previously mentioned, lake level may also fall as a result of a more arid climate. If lake level is lowered without any changes in the paleoslope, the river system will probably need a large time lag to adjust to the new base level, and this will not lead to the same amount of sediment transport as a combination of tectonism and lake level drop. There would be a progradation of the alluvial wedge, but the ratio between fines and sand would probably have been higher.

However, the interaction between tectonics and climatically controlled lake level fluctuations on the potential accommodation of the Colton alluvial system seems complex. The combined effect of climate and tectonics may have resulted in lake level changes of a high temporal frequency and large magnitude. A complicating factor has been the low slope gradient of the southern basin margin, which would have enhanced the spatial effect of such changes, but have dampened the effect on the fluvial system with regards to changes in channel pattern. If the lake level rose as a result of a wetter climate and increased run-off, it is reasonable to expect that this climatic change also gave rise to increased sediment discharge. If sediment supply were sufficient, this may have resulted in the few examples where coarse clastic sediments were transported into the lacustrine environment. The absolute, and even the relative importance of these factors are difficult to establish, but can be suggested.

7.3: Paleocurrent

The paleocurrent data from the SAFARI project are put together by Audun Kjemperud and Edwin Schomakher of UIO and shown in figure 7.4. It shows a general paleocurrent trend of 305° with the bulk of the measurements distributed within a sector of 110° and all within a range of 205. The general paleocurrent trend from the panels 17 and 18 is 330° (figure 7.5). Both calculations show that the Colton Formation was deposited in a fluvial system with a northwesterly trending paleoslope. Too few measurements were taken in panel 17 and 18 to draw any conclusions based on the deviation between the two diagrams. The largest methodical problem is that the paleocurrent rose diagram of panel 17 and 18 is based on both large 3D dunes and to a large part on small scale structures (ripples), and the measurements should be weighted hierarchically. Besides, small scale structures are not recommended for establishing a paleoslope direction, as they are likely to be very much influenced by small, local current directions.

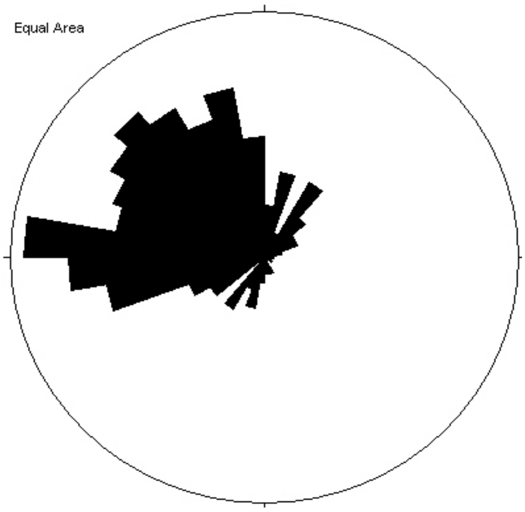


Figure 7.4 Paleocurrent rose diagram based on 351 measurements from the Colton Formation of large 2D and 3D dunes, gutter casts and cut bank orientation. The diagram is an equal area diagram where the outer circle represents 9% of the total population.

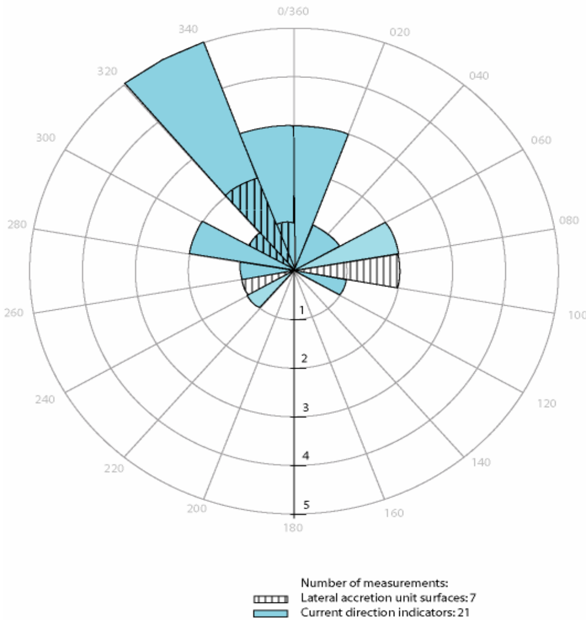


Figure 7.5: Paleocurrent rose diagram based on 21 measurements of 3D dunes, groove casts, current lineations and current ripples from the panels 17 and 18.
All measurements are equally displayed.
In addition to these measurements the rose diagram displays measurements of direction of accretion for lateral accretion units.

7.4: Sinuosity of rivers

Sinuosity and gradient

A change in gradient towards a steeper paleoslope may lead to increased sinuosity. Allen (1984) suggested that the sinuosity of alluvial rivers has a minimum at both large and small slopes and reaches a maximum at intermediate valley slopes. Ackers and Charlton (1975) likewise found by experiments that channels at sufficiently small slopes may become virtually straight with an inner meandering thalweg. Ouchi (1985) performed experimental flume studies, and for a mixed-load system increased slope were found to initially cause increased sinuosity, perhaps meandering, and later transition to a lower-sinuosity system, sinuous or island braided. Based on these results, it might be expected that the Colton Formation were disposed for low-sinuosity channels that may have become more sinuous to adjust to tectonically induced subsidence before the sinuosity again decreased.

Sinuosity and paleocurrent

According to Bluck's (1976) outline in Fig 7.6 below, the paleocurrent directional spread opens the possibility for both an interpretation of the Colton Formation fluvial system as consisting of braided or low-sinuous rivers and more sinuous rivers. The percentage of measurements with a large deviation from the mean is small, but it might reflect some small intervals of meandering systems. The paleocurrent variance should however be used with care, as the preservation potential for structures showing the whole range of flow direction is not equally distributed. Reliability of a sinuosity evaluation from this material is however weak, as the measurements are taken from all levels in the formation, and so it is not taken into consideration that the nature of the stream pattern may have changed vertically.

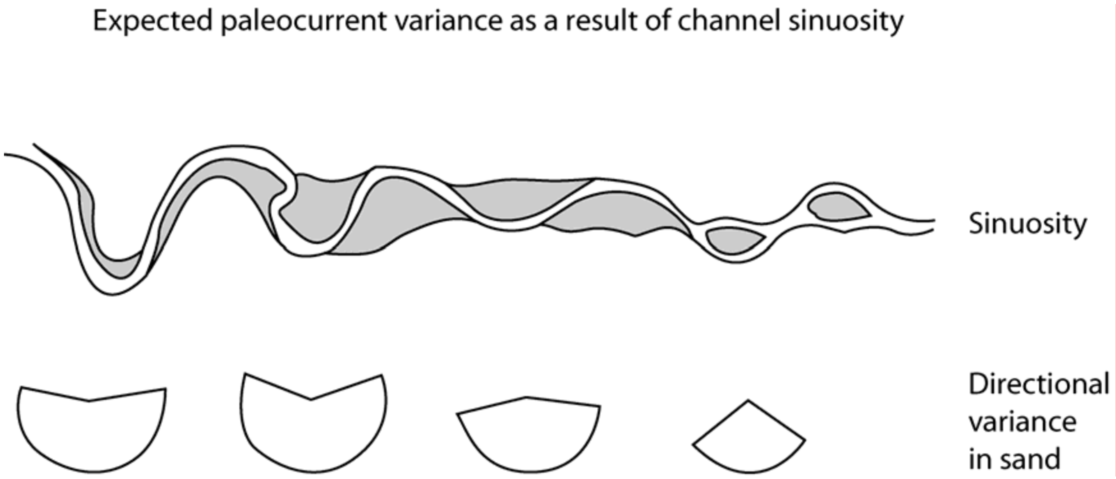


Figure 7.6: Paleocurrent variance as a function of sinuosity (from Bluck, 1976)

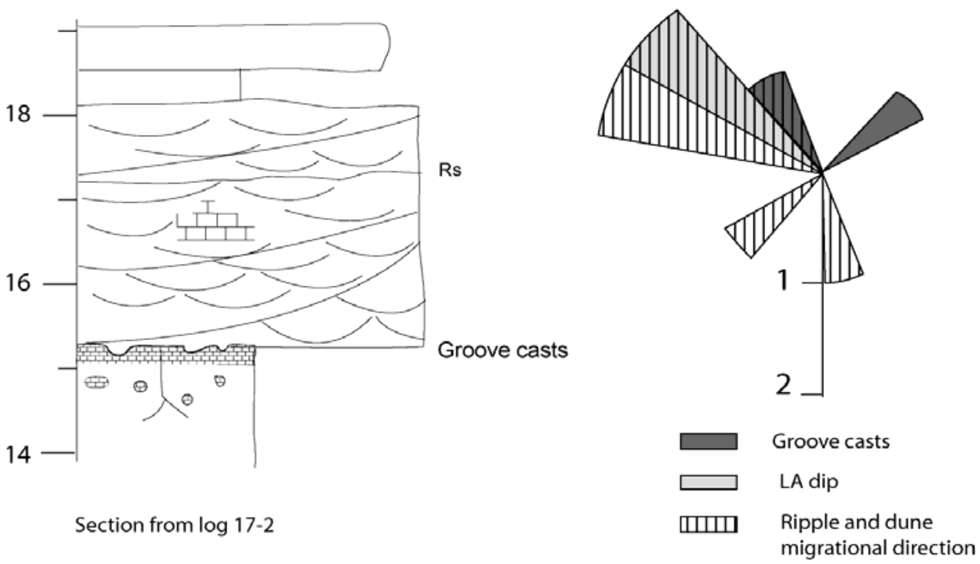


Figure 7.7: Flow pattern in a single channel infill from log 17-2, 15 – 18 meters.

The internal distribution of directions given by the lateral accretionary units may indicate the degree of sinuosity of the channel, as a point bar surface is more likely to have a larger spread of dip angles than lateral accretionary bars and to be almost perpendicular to the flow direction of the channel. Consistent and low spread of the strike of the lateral accretionary elements through a lateral succession supports interpretation of a low sinuosity channel. Interpretations based on the deviation between lateral migration direction and flow direction can however only be performed on one channel unit in the panels, due to lack of measurements, as seen below in Fig. 7.7. The rose diagram in Fig. 6.6 is based on very few measurements. It can still be assumed that the channel is sinuous, since more measurements could not possibly reduce the spread.

Sinuosity and architecture

As mentioned in Chapter 6, the observed channel elements mostly consist of lateral accretionary elements. When these lateral accretionary elements are organized as successive clinothemes and displays unidirectional migration over large distances, it leads to an interpretation of the channels as low-sinuous channels sweeping across the alluvial plain (Harms, 1982). Within the central channel belt, problems with mapping outer channel geometry arise due to extensive reworking. Absence of mid-channel bars does not advocate braided pattern, but sandflats can be found both below and within the central channel belt.

Sinuosity and discharge

Climatic changes, in terms of perennial *versus* ephemeral stream run-off, might have affected the flow pattern. Baker and Penteado-Orellana (1972) suggested that ephemeral streams are more likely to form a braided pattern in environments that otherwise (due to high bank stability) are disposed for meandering flows.

Table 7.1 below is taken from Richards (1996) and shows how changes in discharge and bedload affects aspects of the fluvial system in terms of increase (+) or decrease (-). As stated previously in this thesis, the climate seems to have changed from a humid to dry/seasonal in the Eocene, with an annual discharge decrease during the dry period. Ackers and Charlton (1970) have suggested as a general principle that bankfull flow control meander geometry, and this would imply that average discharge is less important than maximum discharge events. In table 7.1 it is shown that both increased and decreased discharge may lead to a decrease in sinuosity provided that the percentage bedload transport increases. If this is valid, then flood episodes with high discharge (and most likely high percentage of bedload) will affect the channel pattern towards a lower sinuosity. Such fluctuations in discharge can also be expected to i.e. increase the extent of mud drapes and mud chip lags, which could be used to distinguish levels within the stratigraphy of seasonal variability.

Annual discharge	Percentage bedload transport	Channel width	Channel depth	Meander wavelength	Sinuosity	Width to depth ratio	Impact on fluvial system
Increase	Increase	+	+/-	+	-	+	Widening of channels, decrease in sinuosity, increase in slope
Decrease	Decrease	-	+/-	-	+	-	Narrowing and deepening of channels, increasing sinuosity, decreasing slope
Increase	Decrease	+/-	+	+/-	+	-	Narrowing and deepening of channels, increased sinuosity, decreased slope
Decrease	Increase	+/-	-	+/-	-	+	Widening of channels, decrease in sinuosity, increase in slope

Table 7.1: Impact of changes in discharge and bedload (Emery & Myers, 1996)

Mud chip lags and mud drape density within the Colton Formation are largest in the upper half of the central channel belt. The mud chip and drape maxima coincidence with a zone of both relatively deep and shallow channels segments that display both channels with lateral accretion and channels with sandflats. Most of the channels must be considered wider than average, at least wider than the underlying channels, which indicates increased sediment transport. Calcrete is not abundant in the upper part of the channel belt, and most likely the banks were not very cohesive, so the channel flow was not as restricted by the underlying alluvium as elsewhere in the Colton Formation. From the central channel belt to the top of the panels, there is a decrease of mud chip lag on the lateral accretion units in the channels sandstone bodies. These channels are interpreted to have been of low sinuosity, but not as braided. An interpretation could be that these channels are deposited during an interval with perennial high discharge and high bedload percentage.

Floodplain stability, channel w/h and sinuosity

Rather high bank stability can be assumed in most of the Colton Formation since there is high clay content in the extensive floodplain deposits. Further bank stabilization could be by vegetation. Within the channel belt, it is likely that floodplain was less developed or absent in intervals, so that rivers were eroding into sandy, non-cohesive alluvium of former channel deposits. It is generally acknowledged that cohesive floodplain favors increased sinuosity, decrease in meander wavelength and deep, narrow channels (e.g. Allen, 1984). Except for in some intervals towards the top of the panel, grain size of floodplain fines can be considered rather constant and fluctuations of grain size is not thought to have affected floodplain stability.

A low gradient alluvial plain that continues into a low gradient lake margin shelf will not respond dramatically in terms of flow pattern to lake level changes, since the graded stream profile does not need much adjustment to be balanced (Schumm 1994). From the environmental considerations, the sinuosity of the Colton Formation streams is considered to have been low when the paleoslope angle was low, or discharge was high, combined with high bedload percentage, in spite of the rather cohesive alluvium. Periods of tectonic tilting would probably increase the sinuosity temporarily. In non-cohesive alluvium extensive sandflats may have developed, a feature which also would be favored if the flow pattern was extremely seasonal. No true evidence of a braided river pattern is seen, so I suggest that the sandflats are deposited in shallow reaches of straight to low sinuosity rivers.

7.5: Description of intervals in context of varying A/S ratio and sequence stratigraphy

Alluvial stacking pattern is thought to reflect the ratio between the rate of accommodation and the rate of sedimentation in the depositional basin. As discussed above, climatic and tectonic changes in the hinterland may lead to changes in the amount and character of the sediment supplied to the fluvial system that is unrelated to the conditions prevailing in the lower parts. Accommodation is controlled by base level and pseudo base levels, which in turn is controlled by tectonics and climate. To some degree local subsidence due to compaction of underlying sediments may have contributed, but this is not taken into consideration.

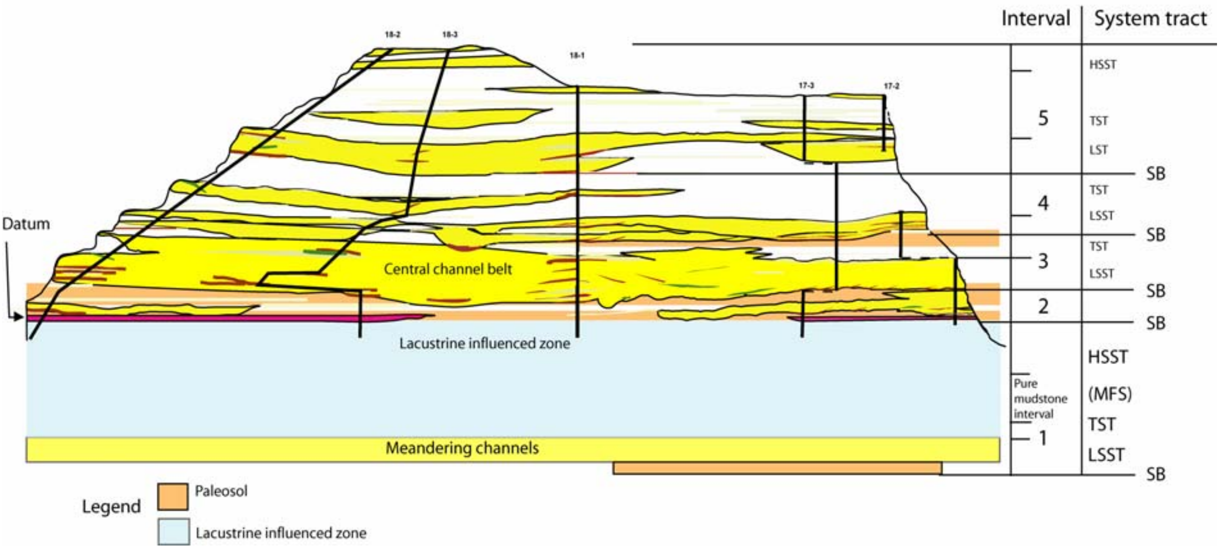


Figure 7.8: Intervals within the Colton Formation

Sequence stratigraphy of this alluvial system is even more afflicted with the difficulties of establishing the degree of influences by the physiography of the hinterland and tectonics and climate. Within these constrains, an analysis of the Colton Formation in terms of sequence stratigraphical systems tracts, stream equilibrium profile and A/S is given below, with reference to Fig. 7.8.

Fossil soils can be used as marker horizons for reconstructing paleosurfaces, but Kraus and Brown (1988) warned against the use of paleosols as correlation surfaces, as lateral development of pedogenesis on a floodplain may be of large and high frequency variations due to a number of factors, as channel proximity, alluvial substratum, drainage pattern and depositional structures.

Interval 1

This is the only interval that can be assigned to a classic sequence stratigraphy since it can be directly linked to lacustrine transgression. It is an upward fining to upward coarsening succession

Lowermost part: Paleosol underlying sinuous channels

The lowermost part of the studied cliff sections consists of grey channel sandstone bodies overlying paleosol, as seen in log 17-1 and in the bottom of log 18-1. Below and above these channel bodies, floodplain mudstone with spread crevasses and most likely distributary channel sandstone occur, as well as some lacustrine mudstone. The underlying mudstone unit is not logged in this study, but belongs to the uppermost part of the Flagstaff Member of the Green River Formation. The erosive surface beneath the lowermost logged sandstone bodies of this study may thus be a sequence boundary.

The paloeocurrent rose diagram in Fig 7.7 indicates that the channel bodies in interval 2 were deposited in a meandering river. If the lake level falls as a result of tectonic subsidence, the increased paleoslope may account for the sinuosity, but in the earliest stages of Lake Uinta the paloeslope must be assumed to have been very slight. Climatic changes towards dryer conditions were more likely to induce a falling lake level in the confined lake, which is supported by the presence of calcrete in the paleosol. Keighley et al. (2003) have proposed a model for the Lake Uinta early stages where sinuous channels are formed as a response to shortening of the stream profile due to raising lake level. This, however, implicates that there was a very short time lag before the river adjusted to the raising lake level.

Middle and upper part: Heterolithic interval with low sand/shale ratio

In the lower part of the panel, above the lowermost channel sandstone bodies, there is a zone formed during a rising base level and transgressive system tract. It is terminated vertically by a laterally extensive beach sandstone body formed by reworking of channel sandstone bodies. Vertical alterations between lacustrine and floodplain mudstones, as well as beach sands (sourced from channel sand and crevasses reworked in a transgressional shoreface) and non-reworked crevasses reflects fluctuations within the general lacustrine transgressive event. Throughout the zone, thin beds of green clay-rich mudstone of likely lacustrine origin alternate with red and brown floodplain mudstone. One may speculate if these small fluctuations within the larger trend can be seen in association with the smallest Milankovitch cycles, they are in any rate most likely linked to climatically induced lake level fluctuations. Base level rise lead to generation of accommodation space and preservation of both fine and coarse grained deposits, seen by the preserved dune forms within the succession, except for in the reworked beach sandstone facies. Sand/shale ratio is low with a minimum (zero) between 50-80 meters below the upper beach deposit, reflecting a period of maximum generation of accommodation space and probably containing the maximum flooding surface. Above these pure mudstones the sandstone percentage increases upwards within the high stand system tract deposits.

Interval 2

Heterolithic interval with paleosol, sinuous channels and sandflat

Just a few meters above the beach sand, isolated channel sandstone bodies are deposited above and eroding into paleosol with calcrete nodules. The paleosol horizon indicates a hiatus or sediment bypass. This can be explained either by a fall in base level with fluvial incision, or that there has been small amount of sediments in the fluvial system. As suggested above, the early stages of hinterland development may have yielded small amounts of sediment. In either case, the paleosol may indicate a dry period, which also may explain a falling base level.

In panel 17, the paleosol indicating the paleohorizon is localized to a surface incised below the beach sand level, supporting a base level fall interpretation and marking a sequence boundary. The channel above also seems to have incised about 15 meters into the fines, following the paleosol horizon. There are isolated or vertically stacked channel sandstone bodies within this short interval, an interval which is ending below yet another paleosol horizon marking the lower boundary of interval 5. Both laterally migrating elements and sandflats are found in this interval. Accommodation seems to be increasing, indicated by preserved channel bodies and levees within the fines (beneath the next paleosol horizon). The paleosol horizon above testifies to the next drop in lake level, most likely climatically induced.

Interval 3

Lower part: Central channel belt, 3a

The central amalgamated channel belt above the paleosol horizon is characterized by a multitude of erosional surfaces within and between partly preserved channel sandstone bodies. Although there are some floodplain fines present, they are terminated laterally by channel floor surfaces. This configuration reflects most likely an extensive reworking of sediments and is caused by a low rate of accommodation space generation compared to sediment supply, characteristic of a low stand system tract in an alluvial setting. Within the channel belt there are alterations between reworked intervals and preserved levees and floodplain fines, reflecting short-scale fluctuations. Although short-scale fluctuations found in lower levels is linked to lake-level fluctuations, the fluctuations within the channel belt are to a larger part interpreted to be a result of changes in discharge and amount of bed load and suspension load sediments in the fluvial system.

Incision can be seen within the belt and it is possible that the central channel belt is deposited in an incised valley. Incision may indicate a lack of accommodation space due to a fall in the lake level, and in addition incision most likely is enhanced by increased discharge. A high sediment supply may have given rise to vertical stacking of incising channels; the sediment supply most likely reflects a wet hinterland climate and possibly tectonic uplift. Tectonic uplift in the hinterland coupled with tectonic subsidence in the basin together can account for a low lake level and increased sediment supply. Fluctuations may be caused by the balancing of the subsidence and the increased water input to the lake due to increased discharge, with the additional moderating effect of the climate within the basin region.

Upper part: Upwards fining heterolithic interval, 3b

At the very top of the central channel belt there are preserved convex upward channel sandstone bodies and abandoned channel infill sandstone bodies. Above these, there is a heterolithic zone of floodplain fines and channel sandstone bodies (Interval 6) with a general fining upward trend. The channels reflect fluvial incision up to 10 meters, but within a zone of a lot of preserved floodplain fines. Below the upper channel sandstone bodies there is another preserved paleosol horizon. Accommodation space seems to have increased slowly from the middle part of the central channel belt to the top of interval 3.

Interval 4

Upwards fining heterolithic interval

The pattern of paleosol below amalgamating sandstone bodies transitional to isolated channel sandstone units and increasing amount of floodplain fines is repeated in the overlying zone; a heterolithic zone with a zone of amalgamated multilateral and multistory channel sandstone bodies with an incision of 15 meters in the lower part (4a) with mostly floodplain fines above (4b). In panel 18 the distinction between interval 3 and 4 is not so clear, the lower bounding paleosol is less well developed and in the central part of panel 18 it is missing entirely, or at least, it is not recorded. Carbonate concretions in the lowermost channel body of interval 4 in the central part of panel 18 may indicate that there have been calcrete nodules and a paleosols horizon in the sediment package that has been eroded. It seems as the lowermost channels in panel 18 became deeper as they migrated (mostly westwards), implying that a short-term increased discharge may have led to a deeper incision, removing the paleosol completely.

Percentage of fines within interval 4 is overall larger than in interval 3, and the lowermost channel sandstone bodies are only partly amalgamated with a lot of intervening floodplain fines preserved. Channel bodies are mostly ribbon bodies, a result of lateral migration. Nevertheless, accommodation space generation relative to sedimentation must have been overall larger than in the underlying interval.

Interval 5

Upwards fining heterolithic interval with isolated channel bodies

This interval shows similarities with the two next underlying intervals, but the overall fines/sand ratio is even larger. The lower boundary has no mature paleosol development with calcrete nodules except for in 18-1, and it is absent in the upper channel bodies, which I interpret to indicate an increased sediment supply on the floodplain. This interpretation is supported by that there are soft sediment deformation in the lower part of this level, a detail which is commented below. An increased depth of lower channel incision may reflect that a possible increased sediment supply is accompanied by increased discharge, which would be expected.

Channel sandstone bodies, channel bedforms (including levees) and crevasses are preserved, indicating an increased accommodation space and steady sediment supply. Crevasses occur frequently within this interval, which can be seen as a predictive tool for the building of alluvial ridges and stable channels banks frequently overtopping. Towards the top of the panel, the succession of interval 5 appears to be upward coarsening, possibly reflecting a reduced rate of accommodation space generation.

An interesting detail is that soft sediment deformation is abundant in the upper part of the Colton Formation within the studies sections of panels 17 and 18 (Fig. 7.9). This may indicate large sediment supply and rapid deposition (Allen, 1984). Soft sediment deformation is also commonly triggered by tectonic movements as earthquakes (Alexander et al., 1987). A large sediment supply might be linked to tectonic movement in the hinterland uplifting the source area.

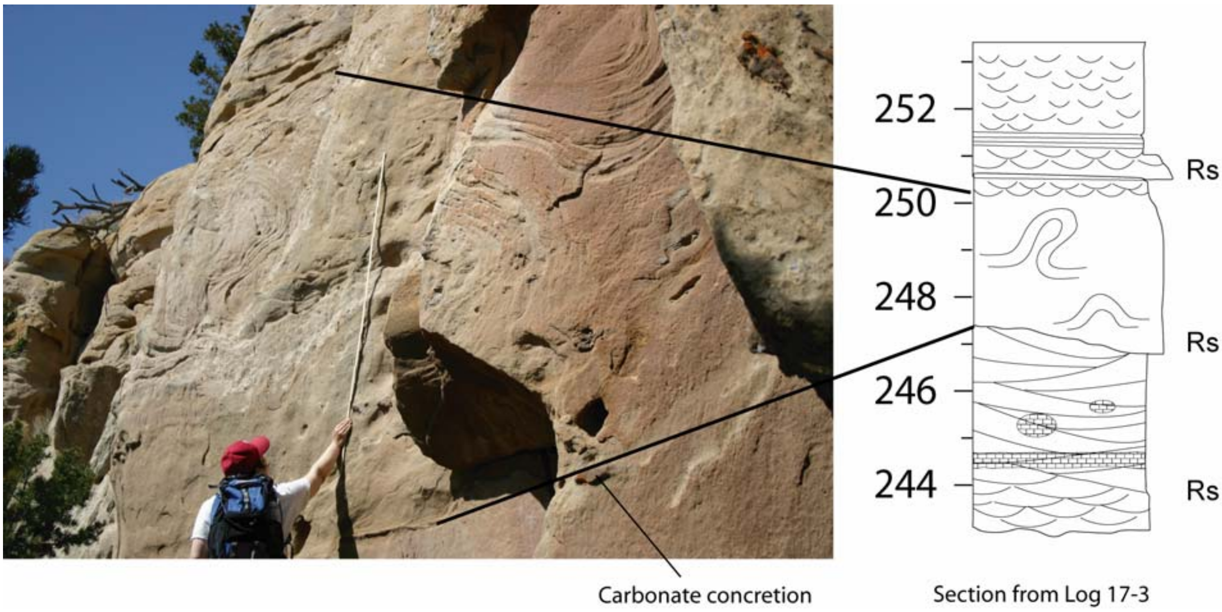


Figure 7.9: Soft sediment deformation

Several factors indicate that this was a tectonically active period, and the increased amount of sediment supply may additionally be explained by the growing maturity of the catchments area. Hinterland uplift and a wetter climate led to increased sediment supply and water input into the lacustrine environment, raising the base level in the distal part of the fluvial system. Intervals 4 and 5 probably represent a more distal position of the alluvial plain relative to that of interval 3, reflected by the difference in sand/fines ratio and possibly triggered by a general increase in lake level that affected an increase in accommodation space in interval 5. Interval 5 is in the SAFARI panels further to the east succeeded by package of distal alluvial plain facies of mudstone and thin and isolated channel sandstone bodies, before the entire Colton Formation is capped by the open lake facies of the Green River Formation.

7.6: Conclusion: Conceptual model of the Colton Formation

The Colton Formation is a fluvial formation, deposited medial and distal on the alluvial plains on the southeastern side of the Uinta Basin. It built out into the deepest part of the basin in the northwest (Fig 7.1) at an oblique angle to the Uinta Basin axis. The rivers advancing over the shallow dipping southern and eastern margin of the Uinta Basin may have been affected by changes in the angle and plunge of the paleoslope brought on by tectonic movements with migration of channel belts and avulsion as the results.

Cohesive banks have led to development of relatively narrow, steep-sided channel trunks regardless of sinuosity or braidedness of the river. The medial alluvial plain sediments consisted of deposits by channels of varying sinuosity and floodplain sediments. The most distal alluvial plain had shallow distributary channels along with fluvial and lacustrine mudflats, having a transition to lacustrine conditions consisting of a mudflat and shoreface environments. The lacustrine environment are reflected in green mudstone and claystone and sandy beach facies formed by reworking of channel sands and crevasses during lacustrine transgressions.

The macro-architecture, comprising the overall alluvial clastic wedge constituting the Colton Formation in the lacustrine host sediments of the Green River Formation and the stacking of channel-fill sandbodies, was tectonically controlled by fault activity leading to basin subsidence and hinterland uplift. Meso-architecture of the Colton Formation deals with the architectural style of fluvial deposits and intercalations of lacustrine and distal alluvial mudflat intervals. This heterogeneity scale was controlled to by both tectonics, climate and base level changes as well as the floodplain stability. This is an expected development going from allocyclic control to a larger degree of autocyclic control.

Colton Formation was initially interfingering with lacustrine sediments (interval 1 and 2), followed by a large two-step drop in lake level, regional channel insicion and renewed rise in base level that culminated in the maximum progradation of fluvial sediments into the basin in interval 3. The retreat of the fluvial system towards the basin margin is found in at least two steps in interval 4 and 5.

8 SAFARI project and well data

Additional data on sandstone body geometries and extent are present in the SAFARI panels. These data are important for establishing better averaged values for channel geometries and channel belt length. All the panels form a continuous diagram of sections parallel, diagonal and transverse to the paleoflow direction for 5.7 kilometers (Fig 8.1), giving an almost three dimensional control.

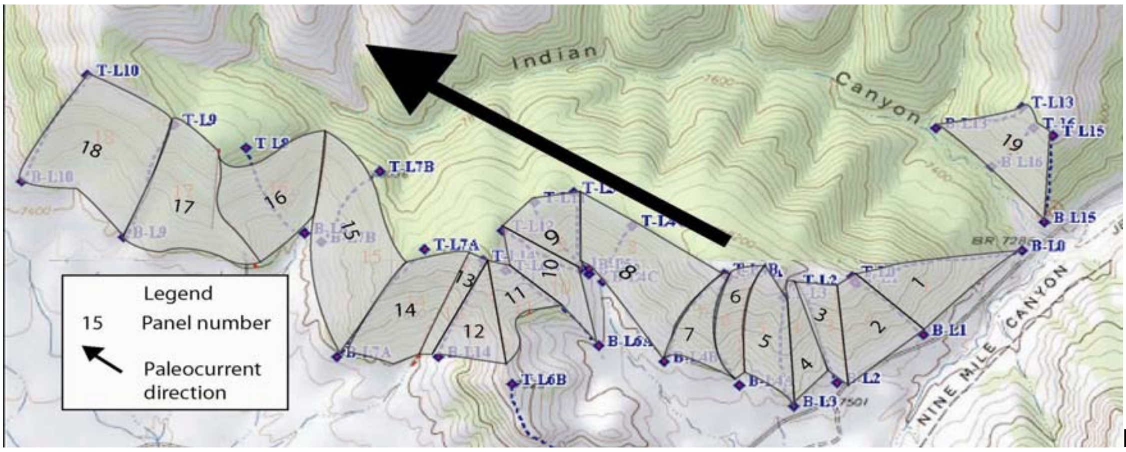


Figure 8.1: All panels in map and paleocurrent direction. North is upwards. Panel 18 and 17 are adjacent to the SAFARI panels in the east.

8.1 Channel belt length

Below is a table showing channel measurements taken from all the panels, including panel 17 and 18. Maximum channel belt width is limited by outcrop width, and maximum width and average width may therefore be underestimated.

Table 8.1: Channel width estimates from the SAFARI project data and W/T estimates. Channel widths are measured where channel margins are observable and calculated using general paleocurrent direction of Colton Formation in the area. All values in meters.

Interval	2	3	4	5
Average channel belt width	254	383	313	190
Maximum channel belt width	379	570	500	440
Minimum channel belt width	49	150	163	85
Average W/T	42	20	19	14

8.2 Channel belt sandstone interconnectedness

Channel sandstone proportions for the SAFARI logs are given below in table 8.2. The proportions are given for both the central channel belt and the vertical interval going from the lower boundary of the central channel belt to top of interval 4a. As it can be seen in the table, central channel belt sandstone proportion is generally higher than the composite interval sandstone proportion.

Table 8.2 a: Percentage of sandstone in central parts of the formation from west to east.

Panel	**17 and 18**		**16**	**15**	**14**	**13, 12 and 11**		**10**
Log	**10**	**9**	**8**	**7b**	**7a**	**14**	**6a**	**12**
Percentage sst in CCB	100	100	80	97	100	100	100	85
Percentage sst in interval of CCB and 4a	68	84	75	97	100	100	93	82

Table 8.2 b: Continuation of table 8.2.a eastwards.

Panel	**9**	**8**		**7**	**5 and 6**	**4, 3, 2 and 1**			
Log	**11**	**15**	**4c**	**4b**	**4a**	**3**	**2**	**1**	**0**
Percentage sst in CCB	100	80	100	100	100	96	83	70	68
Percentage sst in interval of CCB and 4a	83	75	78	75	84	88	80	62	52

The interconnectedness of channel sandstone bodies (c.f. Bridge and Mackey, 1993) from all SAFARI panels and panel 17 and 18 is given below in table 8.3 as sandstone body length connected in outcrop. Interval 2 has a markedly low percentage of length connected compared to total length of bodies along section. This reflects the multilateral as opposed to multistorey architecture of the interval, which is also demonstrated in the low thickness of connected sandstone bodies. The central channel belt sandstone body is dominating interval 3, which has the largest connectedness and thickest connected intervals, and the erosive termination of sandstone bodies can be read from the relatively short average length. The lowest percentage of interconnectedness is found in interval 5, reflecting the high proportion of single channels in this interval.

Table 8.3: Percentage of interconnectedness of channel sandstone bodies along all panels, percentage of interconnected length to the total length of the sandstone bodies, average total thickness of the interconnected sandstone bodies and the average length of sandstone bodies in the panels for intervals. No correction is made for edge effects.

Interval	2	3	4	5
Connected sandstone bodies in outcrop	83 %	100 %	93 %	73 %
Sandstone body length connected in outcrop	38 %	93 %	78 %	54 %
Average thickness of connected sandstone bodies	17 m	34 m	27 m	20 m
Average length of sandstone bodies	287 m	230 m	251 m	214 m

8.4 Comparison of intervals

The lower boundary of the central channel belt sandstone body (interval 3a) is especially distinct as a regional feature. The degree of interconnectedness between interval 3 and 4 is on average larger in the SAFARI panels than found in panels 17 and 18. Incision of the central channel belt sandstone body by the channel erosion surface of the overlying 4a sandstone bodies is up to 15 meters. There is a large variation in the relative connectedness of these intervals. In panels 14 and 19 the two intervals are forming a continuous package of amalgamated sandstones, but in panel 1, 2 and 3 they are completely separated by fines for a lateral distance of 600 meters.

Incision below the central channel belt is up to 20+ meters, in panel 2 almost reaching as low as the datum. The degree of amalgamation within the central channel belt sandstone body varies, reflected in the channel sandstone percentage. The degree of connectedness between sandstone bodies within the zones 3 and 4a is of interest for the formation’s reservoir properties. If the basal beds of the channel belt sandstones are sufficiently permeable, the zones 3 and 4 could be considered as one reservoir unit

8.5 The well data

The position of the well is marked on the Fig. 3.2, but the well is not further identified since the well log data are unpublished.

There is a marked increase on the gamma reading 330 meters down into the log. If shale content decreases abruptly upwards as a result of less lacustrine influence, experience from the outcrop makes it likely that the datum level should be found in the uppermost part or directly above the zone of lacustrine influence. From experience in the outcrops, the top of the central channel belt sandstone unit is approximately 60 meters above the datum, and occurrence of sandy channel units should be prominent at least 200 meters above the datum. The top interval in the well is, however, not necessarily the top of the Colton Formation, as almost 60 meters of fines can occur between channel sandstone units within the upper parts of the formation, as in log 17-2, 17-3 and 18-2.

The lowermost massive sandstone above the high gamma readings in the well is interpreted to be the datum. The top of the Colton Formation is then above the uppermost reliable well log, as data above 70 meters are considered flawed.

9 Heterogeneities

9.1 Introduction

Hetrogeneities in the context of this study deal with differences in permeability between lithologies and the extent and shape of the elements with contrasting permeabilities. General statements about heterogeneity below are primarily sourced from Weber (1986) unless other references are cited.

Sedimentological heterogeneities exist on several levels; megascale, macroscale, mesoscale and microscale (e.g. Etheridge et al., 1998), visualized in Fig. 9.1. Megascale heterogeneities deal with large kilometer-scale contrasts generated by allocyclic control, as basin infill versus underlying sediments or basement. Macroscale heterogeneities are mostly formed by contrasts between average permeabilities in fluvial channel sandstone infill opposed to floodplain or lacustrine mudstone. Macroscale heterogeneity is in the scale of hundreds of meters laterally, and can be of a scale making well-to-well correlation possible in successions where channel sandstone bodies and floodplain mudstone units are stratigraphically well separated. Mesoscale heterogeneity is contrasts in permeability properties within a channel sandstone body consisting of dunes and mud drapes. Determining mesoscale heterogeneity in a reservoir is dependent on models based on patterns found in the well log and/or in a reservoir analogue, as it is in the scale of 10-100 meters laterally. Microscale heterogeneity reflects variation in grain-size, texture, porosity and permeability within and between layers and laminae in a single dune and is in the scale of less than 10 meters down to micrometers.

Heterogeneity below macro level has a large impact on reservoir quality and production. Rapid lateral variations in a fluvial reservoir is to be expected, both in the macro scale, exemplified by an extreme situation when channel sandstone bodies encased in mudstone lack well-to-well communication and on a meso- to

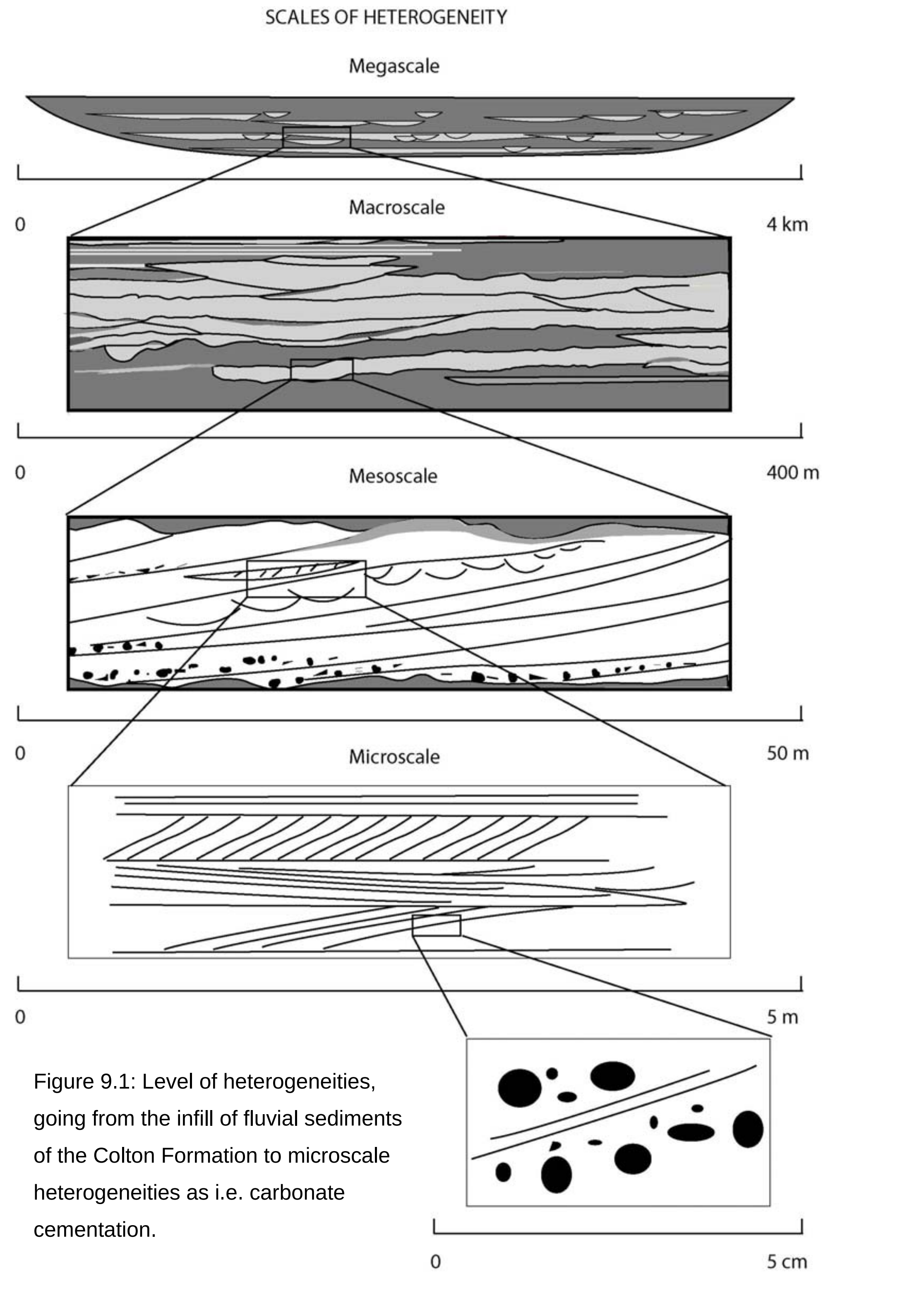


Figure 9.1: Level of heterogeneities, going from the infill of fluvial sediments of the Colton Formation to microscale heterogeneities as i.e. carbonate cementation.

microscale when mud drapes on bounding surfaces within a channel sandstone body affect vertical as well as horizontal permeability.

Quantification and correlation of these heterogeneities can be attempted in outcrop studies by logging and visual estimation of sandstone body interconnectedness for mesoscale heterogeneities and more detailed mapping and lab analysis yielding microscale heterogeneities. Correlation in field over large distances results in a broader picture of how the macroscale heterogeneities are distributed in 3D.

9.2 Megascale heterogeneities

The megascale heterogeneity is the Colton Formation clastic wedge bounded above and below by the Green River Formation.

9.3 Macroscale heterogeneities

Macroscale heterogeneity in the Colton Formation is defined by the contrast between sandstone bodies and floodplain or lacustrine mudstone. Outside the central channel belt unit, the lateral heterogeneity is large. Even when channel sandbodies are in contact due to erosive interfaces, connectivity may be poor because of mud-lined bounding surfaces or because of conglomerates with high clay content deposited above the bounding surface (Fig. 9.2). The data set of the SAFARI is of importance for evaluating the lateral extent of the macroscale heterogeneity, especially heterogeneities concerning the central channel belt.

Sandstone body interconnectedness

Channel sandstone body interconnectedness is dependent on the rate of increase in accommodation as an equally important factor as the type of river system. Shanley & McCabe (1998) stated that independently of the type of fluvial system, low rate in rise of accommodation will lead to extensive reworking of fluvial sediments and a higher preservation potential and channel interconnectedness. In chapter 8, values for the channel sandstone body interconnectedness given.

Bridge and Mackey (1993) performed repeated modeling on connectivity of avulsion controlled channel belt sandstone bodies. They found that if the proportion of modeled sandstone in a sequence was between 0.4 and 0.75, there was a statistical probability of sandbody interconnectedness, which increased with the proportion of sandstone. When the proportion is over 0.75 most channel-belt sandstone bodies will be connected. When the sandstone proportion is below 0.4 sandbody interconnectedness was unlikely.

The percentages of sandstone and fines are given below in table 9.1. Table 9.1 displays the percentages based on the facies percentages from the log, and the depositional architecture is not taken into consideration. In addition, a simplistic macroscale approach is chosen as sandstone percentages are taken solely from the channel sandstones (as Bridge and Mackey's (1993) approach). Overbank facies are regarded to be among the fines, since they are supposed to have so low permeabilities that they could not effectively be a part of a reservoir. Crevasse sandstone beds deposited on floodplains are also counted as fines as they are considered to be a part of the internal heterogeneity within the floodplain fines. As can be seen in the tables, the 17-1 log is not representative for the whole studied stratigraphic section, since it does not extend through the panel, but it is of interest since it reflects the sandstone/fine ratio in the lowermost amalgamated channel sandstone intervals

The sandstone percentage thus varies between 35% and 22% along the section when sandstone data are taken from the log and in addition regarded as possible reservoirs. Lateral separation of the logs are not constant vertically, the separation is commonly decreasing upwards as a result of the topography, and varies between 300 and 40 meters.

Table 9.1: Percentage of sandstone and fines along section based on all facies, percentage of sandstone and fines along section based on channel sandstone opposed to overbank and floodplain deposited sediments.

Percentages	Log	17-1	17-2	17-3	18-1	18-3	18-2
All sandstone and conglomerate facies percentage		71	37	37	36	42	35
Silt and mud percentage		29	63	63	64	58	65
Channel sandstone percentage		60	34	22	28	35	28
Other facies associations percentage		40	66	78	72	65	72

The channel sand proportion in the panels is given below in table 9.2. Interval 3, 4 and 5 seems to have a good possibility for channel sandbody interconnectedness according to the results of Bridge and Mackey (1993).

Table 9.2: Sand/shale ratio based on channel opposed to overbank and floodplain deposited sediments, percentage by intervals.

Log	**Interval**	**2**	**3**	**4**	**5**	**3 & 4**
17-2		0,57	0,48	0,21	0,48	0,34
17-3		0,35	0,43	0,13	0,39	0,25
18-1		0	0,58	0,31	0,5	0,44
18-2		0	0,59	0,46	0,46	0,54
18-3		0,25	0,53	0,26	0,41	0,36

Taken from the panels, the macro scale heterogeneity can be established by estimating the lateral extent of connecting sandbodies. The lower part of the intervals defined in chapter 8 all have the best lateral connectivity, which is to be expected as the intervals are defined by a low A/S at the lower boundary (as discussed in Depositional environment). Interval 3, the central channel belt has a clear high interconnectedness of individual channel sandstone bodies in the panel, 4 and 5 for about 600 meters, interval 2 has poor interconnectedness and in interval 1 is the interconnectedness is not established, but assumed to be good. Thicknesses of these intervals are given in the table 9.3 below. Vertical heterogeneity varies within shorter intervals than the lateral heterogeneity on all scales in the fluvial system; layering, reflecting the nature of the fluvial depositional environment.

Table 9.3: Thickness of multilateral sandbodies in panel 17 and 18 by intervals.

Interval	1	2	3	4	5
Thickness of multilateral sandbodies	20 m	none	24 – 48m	15 m	10 - 28 m

Previously in this chapter, crevasses and overbank deposits were discarded as potential reservoirs. However, some of them have high sandstone percentages, and when crevasses form as avulsion sediments (see Facies associations) they may be deposited close to channel sandbodies (as seen in Fig. 9.3). If there are no mud drapes present, channel sandstone bodies separated by overbank, thin layers of floodplain and crevasses may be communicating, although permeability in the zone between the channel sandstone bodies will be reduced. A crucial factor in a reservoir will then be the viscosity of the hydrocarbons present, as low-viscosity hydrocarbons will be able to be drained or migrate in these zones of lower permeability. Individual estimates of the permeability contrasts between channel sandstone and the other deposits must be made, as hydrocarbons will not migrate through the intervals if the contrast is too large

Høimyr et al. (1993) performed a modeling study on a flow and sweep efficiency on the middle Statfjord Formation. The study compared three models of the same averaged mean permeability. The mean effective permeability of a model with homogenous permeability, lateral permeability zonation and heterogeneity discontinuity obtained from a braided stream analogue formation in Colorado. The results showed a large impact on the flow and sweep efficiency of the heterogeneities from the reservoir analogue was large for the analogue model, having an up to 42 % reduction in efficiency.

Figure 9.2: Panel 15 of the SAFARI panels (figure 3.2) Height of section approximately 130 meters. The section has high sandstone body connectivity, but also mud drapes and channel lag conglomerates on the bounding surfaces restricting permeability.

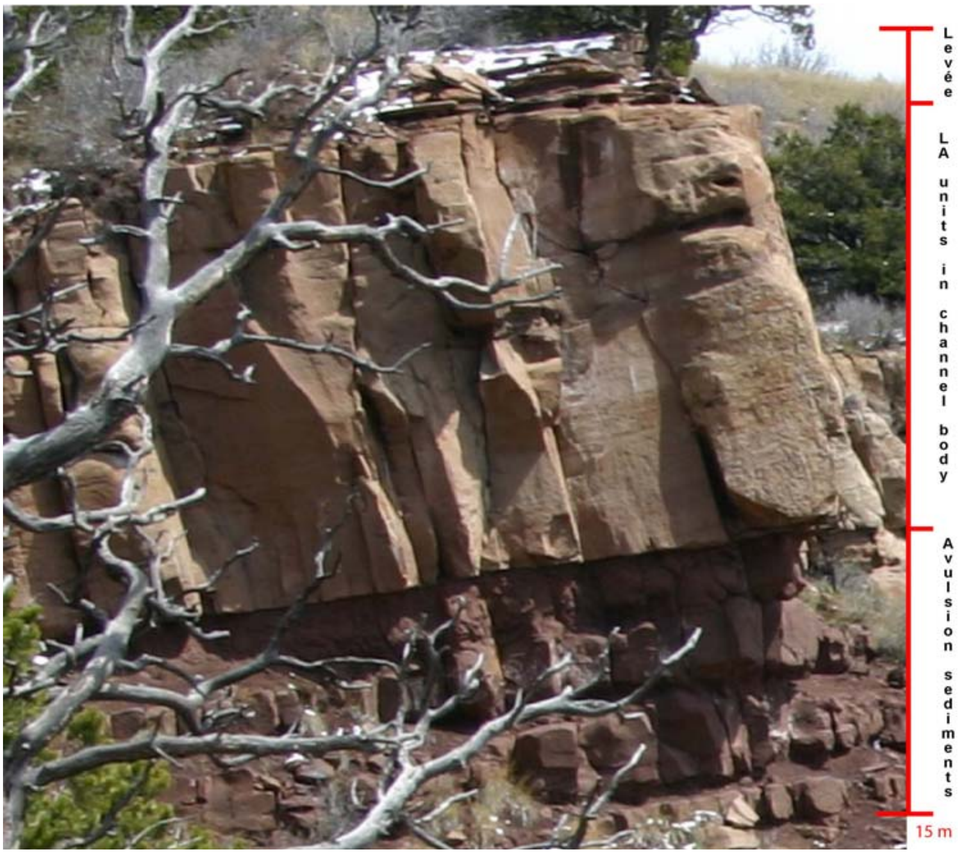


Figure 9.3: Picture from panel 17, showing top of interval 3b and channel sandstone in bottom of interval 4a. This section is not in any log, but is within channel sandstone 22. Note scarcity of conglomerate in the LA units, giving this sandstone package a high permeability in the central parts.

9.4 Mesoscale heterogeneity

Generally, the mesoscale heterogeneities in channel sandstone bodies are mudstone deposits or remnants, either as mud drapes, conglomerates or abandoned channel infill. Internal variations are also found as variations in sediment structure or grain size distribution.

Mudstone drapes can be seen to separate channel sandstone bodies, as in panel 17 where the lower boundary of channel 20 deposits has a mudstone drape. The mud drapes vary from less than 1 cm to 20 cm in thickness, and are most frequently occurring in the upper half of the central channel belt sandstone body and in the upper channel units. The mudstone drapes are laterally discontinuous, usually being terminated within a few meters, but more enduring mud drapes have been found in the central channel belt body, as below channel sandstone 20 in panel 17 (Fig. 3.3).

Figure 9.4 below shows the heterogeneity caused by the thick conglomerate beds in panel 17. Channel basal lag conglomerate clasts are derived from paleosol, and mudstones are likely to be part of the matrix as well as the clast material. Lower channel boundaries, reactivated surfaces and scours between clinothemes are assumed to be permeability barriers, as mud drapes and conglomerates containing mud clasts are found on these surfaces. Mud drapes are not found on single dunes within the clinothemes except in the uppermost part where there is a transition to levees. Sandstone dune beds in sandflats may have mud drapes or mud clast aggradations on top

From the logs it is found that the average vertical distance between these permeability barriers in the most interconnected channel sandstone units, the central channel belt sandstone and the overlying interval 4A are between 1.7 and 10 meters, average value for all logs are 3.9 meters, see table 9.4.

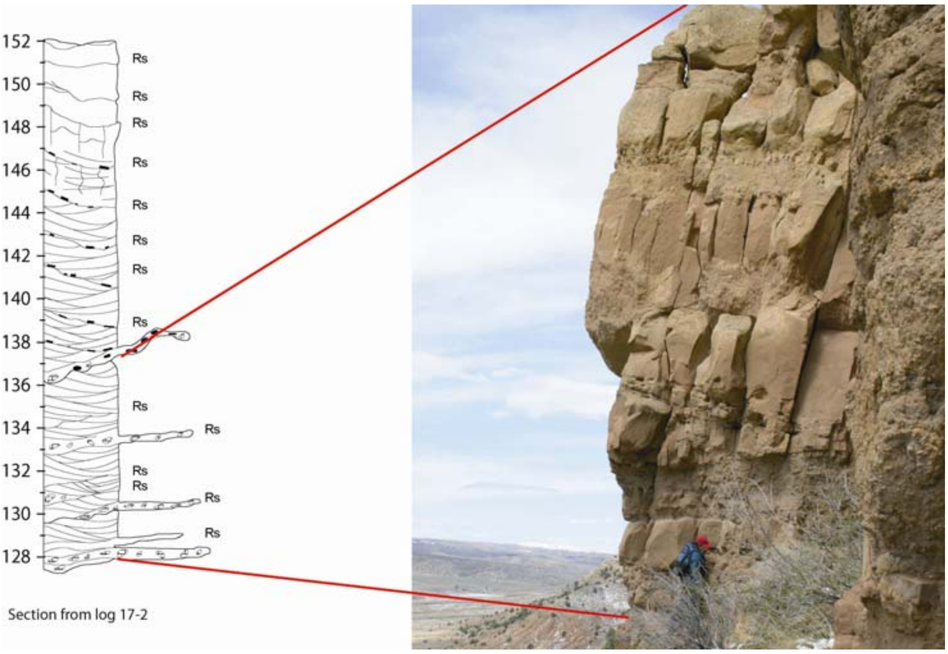


Fig. 9.4: Mesoscale heterogeneity in channel body due to conglomerates in panel 17.

Table 9.4: Number of, and average vertical distance between barriers to fluid flow from the logs along outcrop from east to west. *Floodplain fines between channel sandstone bodies in interval **Mud drapes and conglomerates

	Central channel belt					4A				
	17-2	**17-3**	**18-1**	**18-3**	**18-2**	**17-2**	**17-3**	**18-1***	**18-3***	**18-2***
Number of barriers**	6	14	13	11	8	4	6	4	2	6
Average distance between barriers (m)	4,67	1,71	3,15	4,36	3,38	3,25	1,67	3,25	10,00	3,17

Grain size variations within a single dune occasionally occur (mostly in sandflats), but some of the channel sandstone bodies only show a general decrease in grains size upwards. From table 9.5 below, it can be drawn as a broad conclusion that outside the channel belt, percentages of upwards fining channel successions in the total number of channel deposits fluctuate about 20%. Within the central channel belt sandstone unit preserved individual channel deposits can be of uniform grain size and even levees may consist of sandstone, but the upwards fining channels constitute 57% of the total, markedly higher than in the other intervals. Taking the gamma readings from the well log into consideration, it still seems as if most channel units are distinctly upwards fining or have an increasing percentage of fines, even if it was not recorded in the field. Grain size variations are not necessarily permeability barriers, but should be accounted for as an important aspect in planning procedures in draining a fluvial reservoir. During production, low-permeability zones may be undrained if the production rate is high, giving rise to “thief zones” in the most permeable part of the channel sandstone units.

Table 9.5: Percentages of upwards fining channel sandstone units of total number of channel sandstone bodies in intervals of panels 17 and 18.

Interval	Percentage of upwards fining channel sandstone units
5	24
4	15
3	57
2	20
1	23

The pattern of vertical grain size and facies distribution seems to indicate that channel units have the best permeability in the middle parts, where there is less frequent compartmentalization by conglomerate layers or mud drapes and where the most uniform grain distribution as well as the largest grain size is present.

It is not likely that the laterally accretion (LA) units should be regarded as isolated reservoir compartments, even though both mud drapes and conglomerates may put restrictions on permeability connectivity. and may behave as permeability barriers on a local scale in the reservoir. Hartkamp-Bakker and Donselaar (1993) have found in studies of point bars in Loranca Basin that preserved channel abandonment layers (as mud drapes) may cause partial compartmentalization in a reservoir. The permeability barriers in the Colton Formation are discontinuous due to erosion. Even though mudstone barriers may create tortuous flow paths, they are not likely to be barriers on the whole interface. In addition, the barriers can at places be penetrated by burrowings and the clay content in the conglomerate matrix may be small, so that the basal conglomerate lag is instead a high-permeability zone. An example of sand-on-sand contacts between channel bodies are also seen, as in log 18-1, 185 meters.

9.5 Microscale heterogeneity

When studying the heterogeneities in the Colton Formation, microscale heterogeneities were mostly disregarded. Since the formation is assumed to be a reservoir analogue for architecture and reservoir extent, the specific geochemical or small scale textural composition of the formation is not of primary interest. It could however be argued that the microscale heterogeneities to some degree reflect the catchments area, and that the nature of the sediments on its part holds a strong control on the macro- and mesoscale architecture due to the nature of the clastic input. An example of this is the percentage of fines in the sediments which affects floodplain stability.

A very general assumption can be made on the microscale heterogeneity of the Colton Formation, as it is classified as a feldspathic wacke (see Geological setting). Fluvial sedimentary rocks with a certain amount of clay content generally displays better lateral than vertical permeabilities. This is attributed to both micro- and mesoscale orientation of clay particles both during deposition and compaction (e.g. Clennell et al. 1999).

Also, heterogeneity due to cementation is seen within the formation, and this I previously tentatively linked to initial sedimentological structures, since carbonate cement could possibly be sourced from dissolved calcrete nodules (see Deposional environment).

10 Modeling

Stochastic modeling is performed by randomly populating inter-well areas in a model with geometrical representations of sandstone bodies. Input data from wells form the basis for the sandstone bodies. The inter-well area is populated according to basic geological principles of super-positioning and geostatistical probabilities (e.g. Buller et al., 2003).

According to Weber (1986), 'reservoir heterogeneities should be quantified to a level sufficient to design a reservoir model from which the optimal production process, drainage, and completion pattern can be derived'. As such quantifiable data are W/T ratio of sediment bodies, interconnectedness of the reservoir units, internal facies distribution, grain size and sorting, degree of bioturbation, mud drape length and distribution, diagenesis and burial/decompaction history. Bryant and Flint (1993) emphasized the importance of that geological units to be modeled represent 'flow units' (*sensu* Hearn et al., 1984) as well as genetic units, but commented also that these units are often corresponding.

The aim of the modeling in this study is to generate a principal facies model showing the macroscale channel belt sandstone body correlation of the Colton Formation. Primary macroscale sandstone body correlation is performed for simplicity in this study. General parameters required for the macroscale model is channel belt width and thickness, sandstone percentages, sinuosity and a general paleoflow direction. These values have been presented in previous chapters and are summarized in table 10.1 below.

The modeling performed is a stochastic model with fixed sandstone percentages in vertical intervals between interpreted surfaces. Sandstone body generation during a run is conditioned to the pseudowells, so that logged sandstone bodies will fix the vertical position of sandstone bodies in the modeled outcrop. It is, however, not achievable, or even necessary or wanted, to generate a perfect replica of the known outcrop. Modeling in this study rather provides a tool to experiment with which factors that affects the macroscale heterogeneity of the fluvial deposits.

The Colton model can doubtlessly be improved by refining and increasing the number of surfaces, but this will only serve to lock the model closer to the logged outcrop, and yields little in the way of understanding the governing mechanisms for facies distribution. Establishing trends in parameters as sandstone percentage for larger intervals both vertically and laterally is a more time-consuming, but probably better approach. These trends could reflect A/S variation, and thus have a general applicability in a principal model, since the A/S ratio affected connectedness within the reservoir.

10.1 Colton model

Logged sections were used as pseudowells in the modeling tool. As well as the sections logged in this study, Panel 18 logged by Katrine Brinck (2005), logs from the SAFARI project and the unpublished well data from Minnimaud Canyon were applied in the modeling work. The resulting triangular modeled grid block has 5 km long sides and a thickness varying between 360 and 120 meters. Map outline of the model is shown in Fig. 3.2.

Sandstone facies within channel belts are used in the input logs, as well as the datum beach sandstone for simplifying correlation, and the rest is modeled as floodplain fines. Lateral correlation of pseudowells is based on the datum, and additional surfaces are established below and above amalgamated channel sandstones.

Lower and upper boundaries of the logs were drawn and constructed as surface objects to form the upper and lower boundary of the model grid. Additional surfaces constructed to be used when making the grid and populating the model with zones were the upper and lower boundary of the datum, the lower and upper boundary of the central channel belt unit (3a), and the lower and upper boundary of the amalgamated channel sandstone belt above the central channel belt, 4a. These zones were chosen for simplicity, as these channel belts sandstone unit also amalgamates in the scope of all the panels, and together forms the zone in the Colton Formation with the highest sandstone body connectedness

Within the constructed model grid based on these surfaces additional layering was established, setting cell height. Since mesoscale heterogeneity is disregarded, the cell height can be set to minimum channel height in the intervals. The logs were then up-scaled to the cell size using a simple majority of facies within cells function to populate the facies within the synthetic up-scaled logs.

When the logs are up-scaled, the facies modeling can be performed with channel bodies conditioned to the up-scaled logs. Before the model run, channel geometry parameters for width, height, amplitude, wavelength and orientation were set. These parameters are set with a minimum, maximum and mean value with a factor for drift between these numbers ranging from 0-1 to allow for different degrees of variation within these boundaries. Below is a table (table 10.1) of the parameters used in the various intervals based on facies distribution in the logs and facies association as well as depositional environment analysis performed in the previous chapters.

Table 10.1: Parameters for models.

Interval	1	2	3	4	5
Average paleocurrent direction	305°	305°	305°	305°	305°
W/T of channel belts	24	42	20	19	14
Maximum channel belt width	495	379	590	570	440
Minimum channel belt width	112	49	163	150	85
Average channel belt width	285	254	383	313	190

Run with these parameters resulted in the following models in figures 10.1 to 10.6. The figures are displayed as the outcrops of the SAFARI panels. Fines is green colored, channel sandstone is yellow, and the datum is marked with red. As can be seen from the models, the central channel belt is recognizable as a regional feature above the datum through the whole extent of the Colton Formation, even if its internal sandstone percentage is varying. Also, the upper and lower part of the formation in outcrop has a markedly lower number of channel sandstone bodies. Even if there is a superficial difference between the realizations, there did not seem to be any principal differences between the models based on the sinuosity.

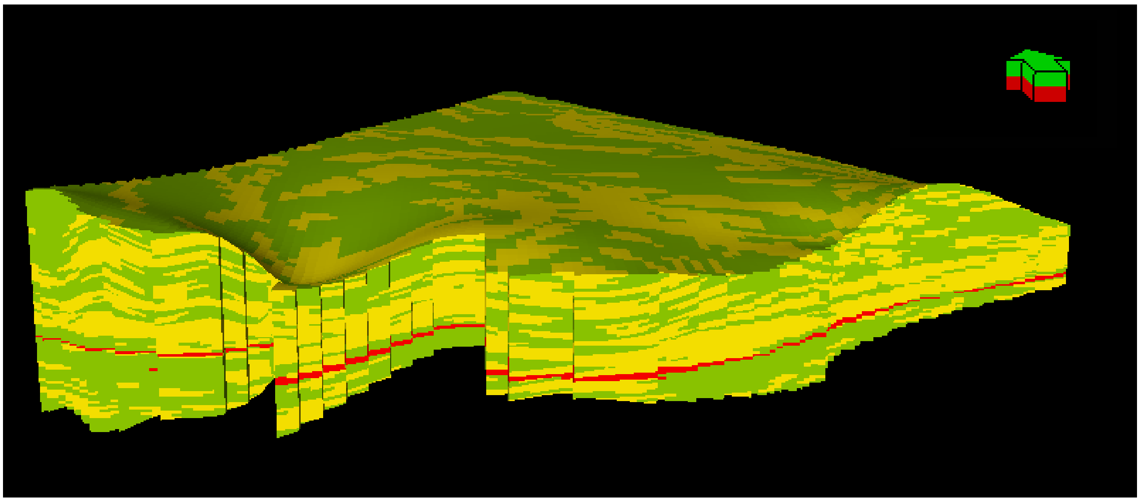


Figure 10.1: Example of facies model realization with low sinuosity. Length of panel 5.7 kilometers from east to west. Three times vertical exaggeration.

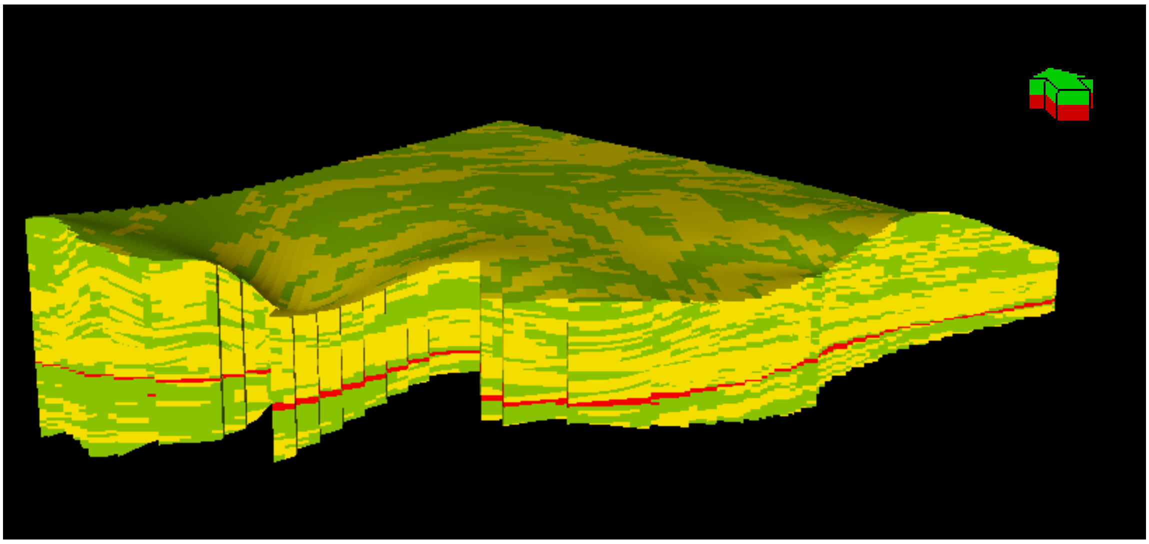


Figure 10.1: Example of facies model realization with high sinuosity. Length of panel 5.7 kilometers from east to west. Three times vertical exaggeration.

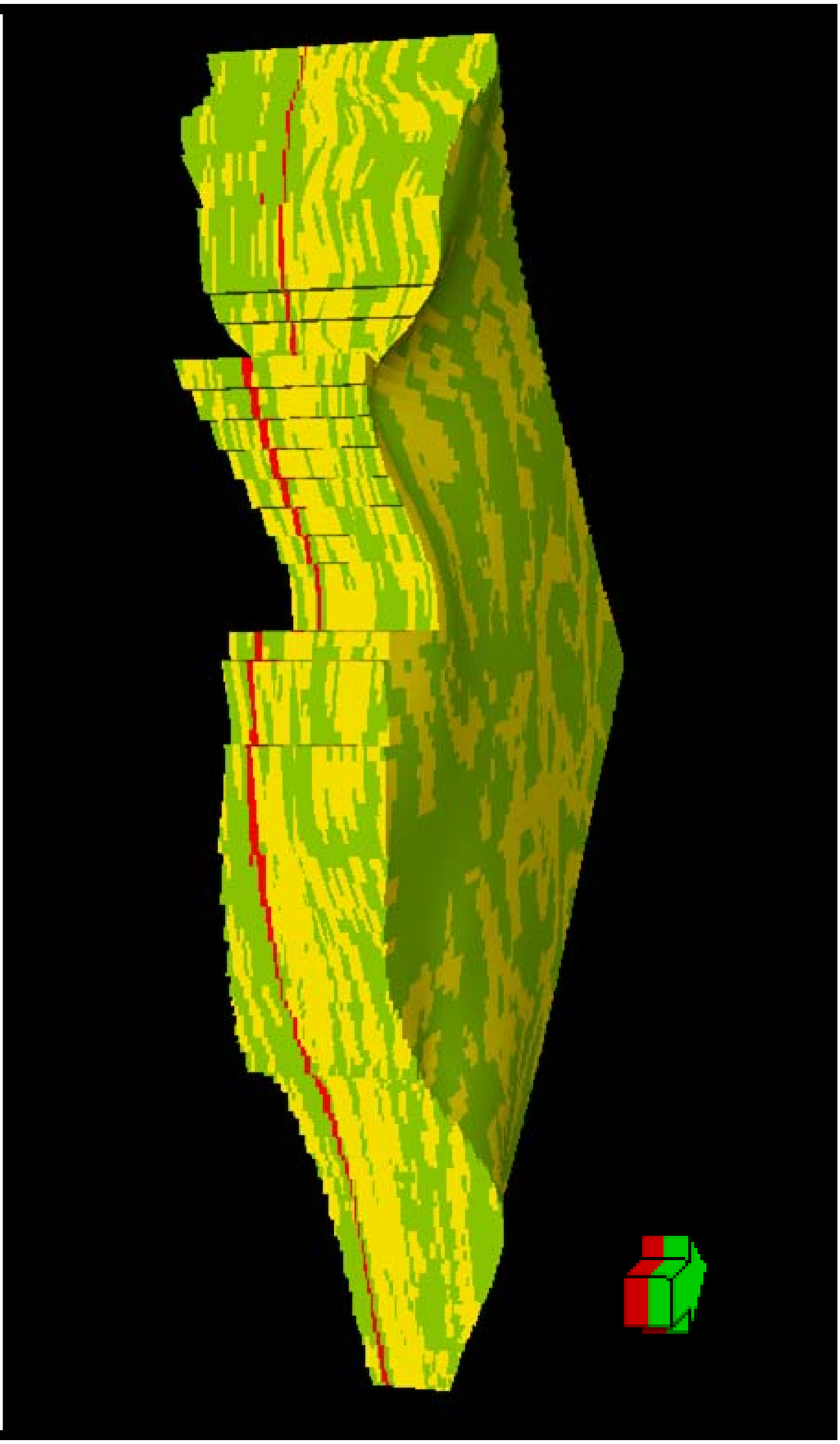


Figure 11.3: Intermediate sinuosity model of the Colton Formation outcrops from the Safari Panels and panel 17 and 18. Extent of the profile is 5.7 km along outcrop from east to west. Vertical exaggeration is 3X.

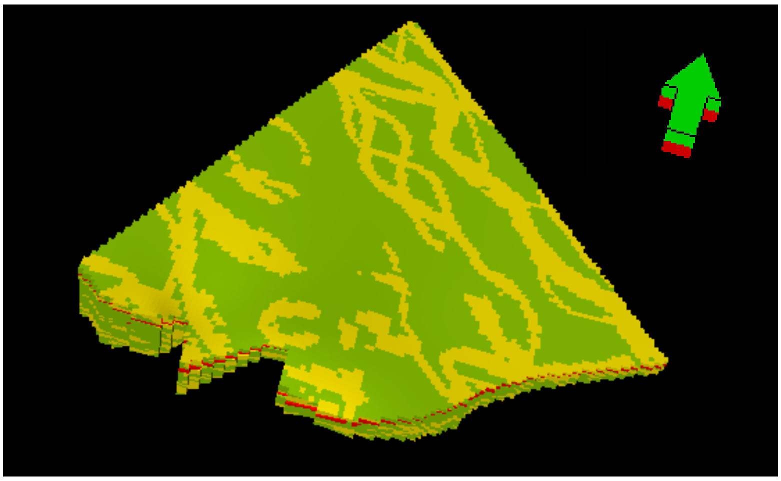


Figure 10.5: Example of surface of interval 2 with low sinuosity.

Figures displayed in the map perspective from figure 3.2. SAFARI panel outcrops to the south and well data as the northern point

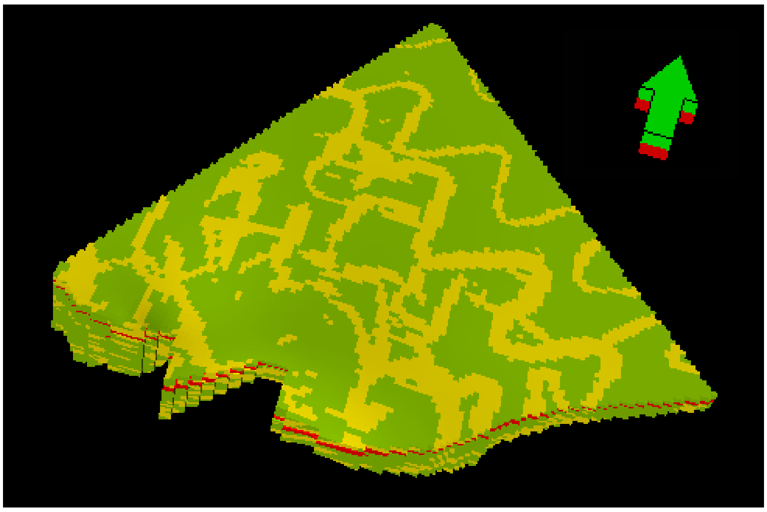


Figure 10.4: Example of surface of interval 2 with intermediate sinuosity.

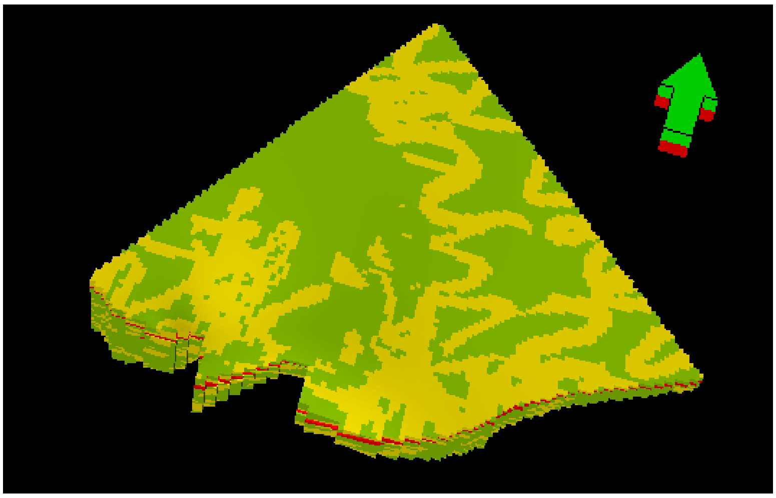


Figure 10.6: Example of surface of interval 2 with high sinuosity.

Below are figures of the modeled sandstone body interconnectedness in Fig. 10.7-10.9. Bodies of the same color are interconnected. Perspective of model is seen from below to display the markedly lower interconnectedness in interval 1. Interconnectedness of the sandstone bodies is high, as also could be expected by Bridge and Mackey's (1993) results. The datum sandstone unit was not included in the interconnectedness model, and would probably increase the interconnectedness even more. Pink, interconnected units are largely above datum. The lowest degree of interconnectedness is found in interval 1.

It must be noted that this is a valid interconnectedness only for fictional homogenous sandstone, since the model ignores permeability barriers on boundaries between channel sandstone bodies as well as mesoscale heterogeneities. The sandstone interconnectedness can not be used as a template for estimating connectivity of a potential reservoir.

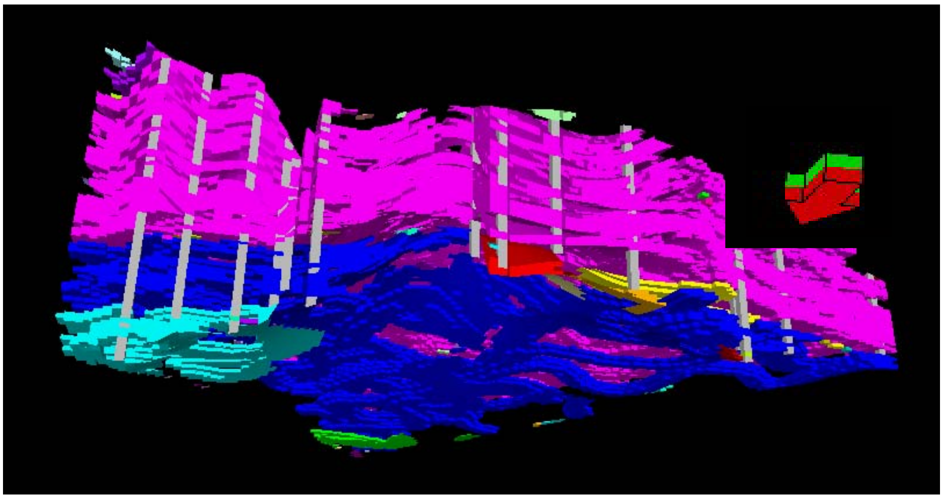


Figure 10.7: Channel sandbody connectedness in facies model realization with intermediate sinuosity. Interconnectedness visualized by color. Five times vertical exaggeration. Perspective from below.

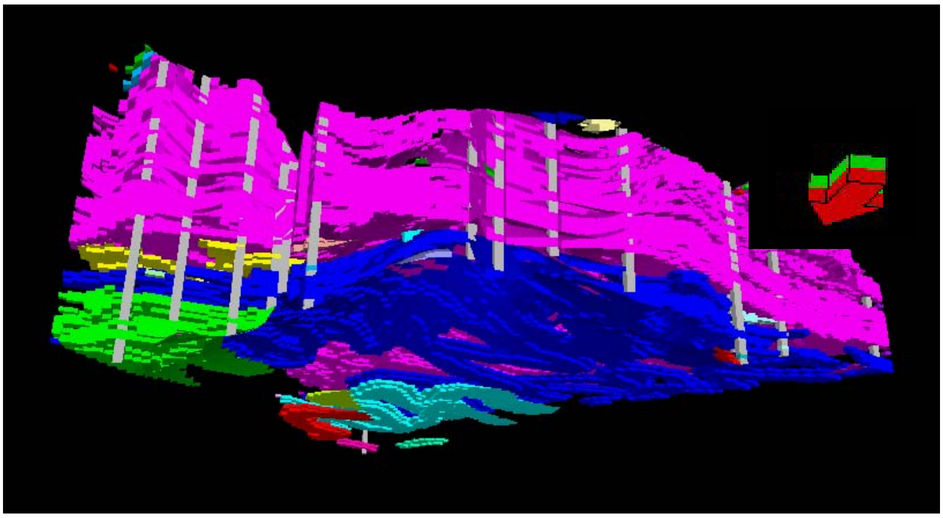


Figure 10.8: Channel sandbody connectedness in facies model realization with high sinuosity. Interconnectedness visualized by color.

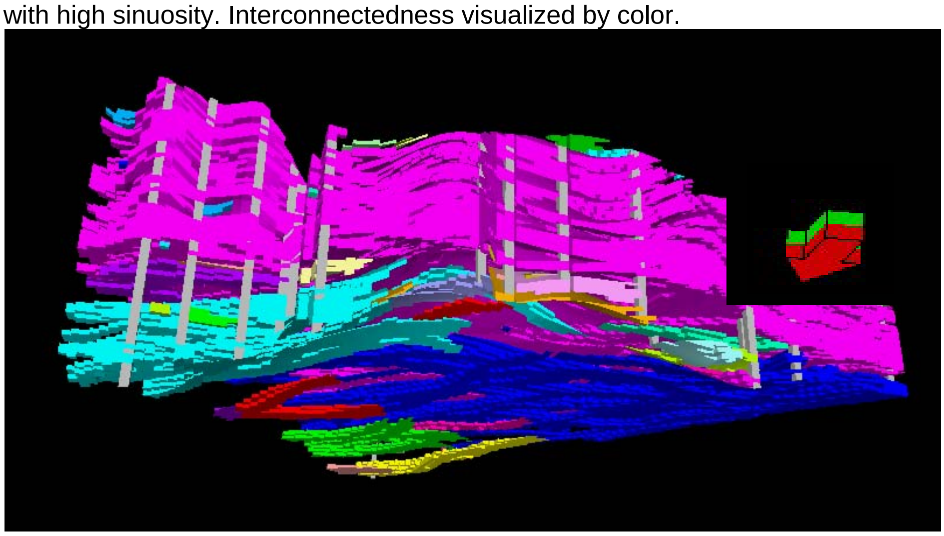


Figure 10.9: Channel sandbody connectedness in facies model realization with low sinuosity.

11 Field analogues in reservoir modeling: Discussion

As mentioned in the modeling chapter, analogue reservoir modeling requires quantifiable data from recent sedimentary systems, wells or outcrop. Quantification from field data is afflicted with problems of interpretation and correct field measurements. In addition, fluvial system has clear directionally controlled proportions on architectural elements. Because of this, lengths of architectural elements in outcrop can be useless without paleocurrent data. Emphasis must also be put into determining the extent of the heterogeneities as for instance mud drapes.

Choice of outcrop analogues is a debated topic. The issue is how to define criteria for determining the choice of analogue outcrop: be it structural depositional setting, sequence stratigraphy or subsidence rates. Within the span of this study, no such analysis is attempted. A more basal problem to be addressed before identifying a reservoir analogue is determining for the specific formation which factors have influenced the depositional pattern and heterogeneity.

The model constructed in this study is very simplistic, and could doubtlessly be improved by more careful analysis of the parameter data, not to mention a more extensive knowledge of the tool. Focus on the modeling of this study has been a coarse division based on segments constructed by A/S analysis. It is one of several possible approaches, supported by Bryant and Flints (1993) statement: ‘It is important to compare like-with-like when compiling data sets, thus sequence stratigraphy provides an important framework within which to group geometrical data.’ This is an approach that seems to work on the macroscale level in a reservoir. Meso- and microscale heterogeneities are more dependent on autocyclic processes. Analogues for these features should perhaps be chosen with more care with respect to similarities in grain size and other features local to the specific alluvial plain.

12 Conclusion

- The Eocene Colton Formation was deposited as a clastic alluvial wedge into the Lake Uinta Basin at the early stage of the lacustrine setting. Rapid fluctuations in the lake level led to alternating fluvial and lacustrine sediments in the lower part of the formation, which reflects the distal alluvial depositional environment. Rocks higher in the stratigraphy of the studied outcrop formed when sediments were deposited on the medial alluvial plain. They show a larger scale cyclicity of repeated sequences of alluvial sediments going from amalgamated channel sandstone belts above an erosional surface to isolated single channel sandstones in floodplain fines topped by the next erosive boundary. Alluvial sedimentary rocks show some evidence of having been deposited subsequent to fluvial incisions over 20 meters deep, especially notable in the Central channel belt. The upward fining cyclicity reflects a generally rising A/S that created increasing accommodation space. Macroheterogeneity of the Colton Formation is formed by this pattern.

- Seasonal precipitation pattern in climate in the early Eocene led to pulses in discharge and sediment supply. Alluvial architecture of low-sinuosity rivers in the upper part of the outcrop is formed during ephemeral bankfull flow with a high percentage of bedload. Rapid deposition of bedload is reflected by soft sediment deformation of channel infill. Seasonal drought is also reflected in the mesoscale heterogeneity by mudstone drapes; mud chip lags and mudstone/calcrete nodule conglomerate on lateral accretion units and as channel lags. These sedimentary structures participate in making the complex mesoheterogeneity of the channel sandstone bodies in the Colton Formation.

- Episodes of tectonic activity leading to large basin subsidence in the north led to a preferred northwards channel belt migration, deflection and also increased paleoslope. Increased paleoslope resulted in increased sinuosity of rivers. Subsequent erosion and deposition due to changes in the stream equilibrium profile with time again lowered the sinuosity of the rivers.

- Reservoir modeling showed that the theoretical sandstone body connectedness in the middle part of the formation is good, especially in the central channel belt unit. Outside the central channel belt, lateral and vertical variation in sandstone percentage makes the potential reservoir complex in three dimensions. The complex mesoheterogeneity will also have a large impact on the reservoir properties. Erosional scouring of low-permeability zones largely prevents isolation of channel elements within the amalgamated sandstones, but fluid flow in the reservoir will be tortuous on both meso- and macroscale.

- The stocastic modeling of the macroheterogenty og the Colton Formation can only be regarded as a model of the potential sandstone body connectedness. To perform a modeling of the reservoir connectivity, mesoscale heterogeneity modeling is crucial.

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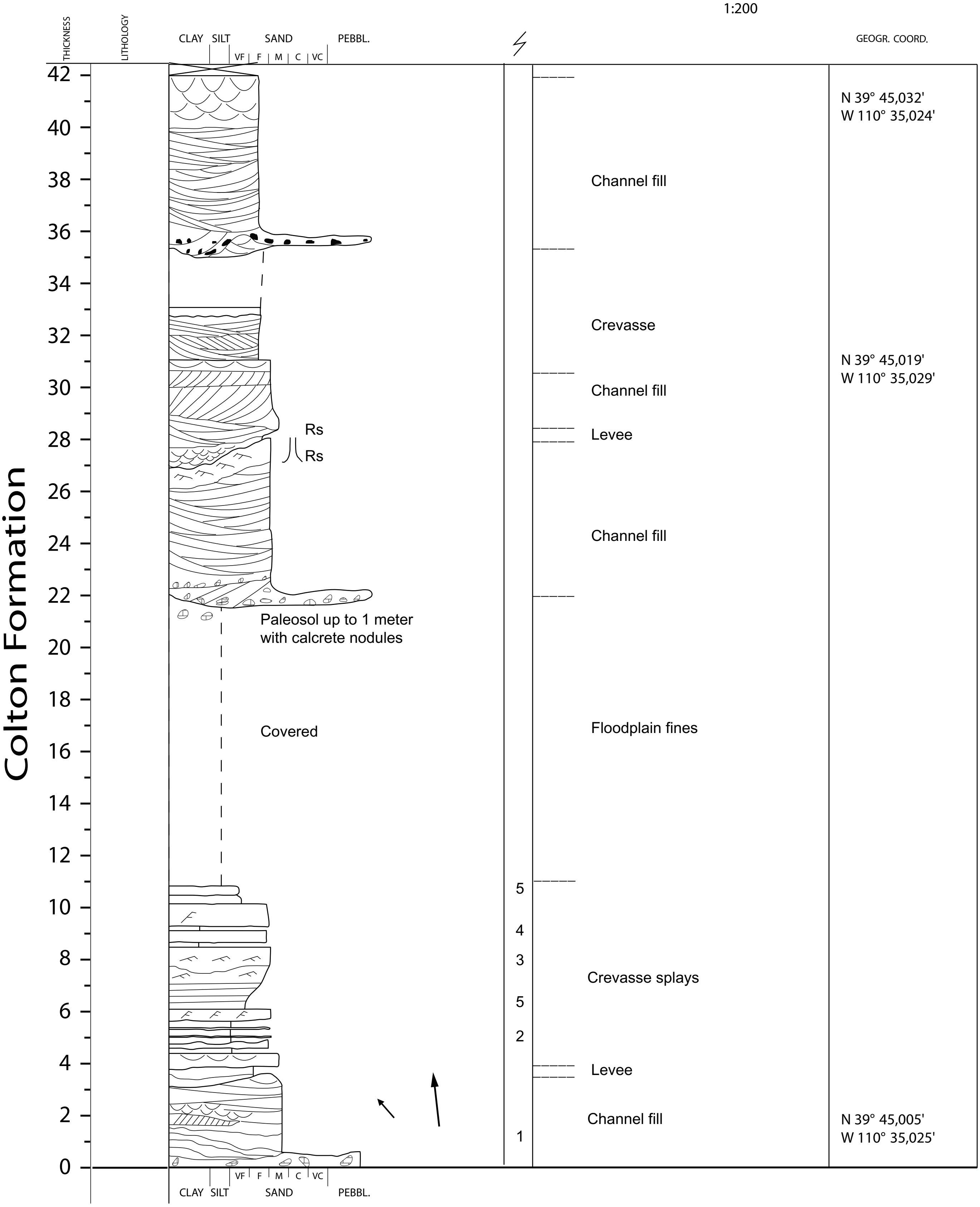
14 Appendix

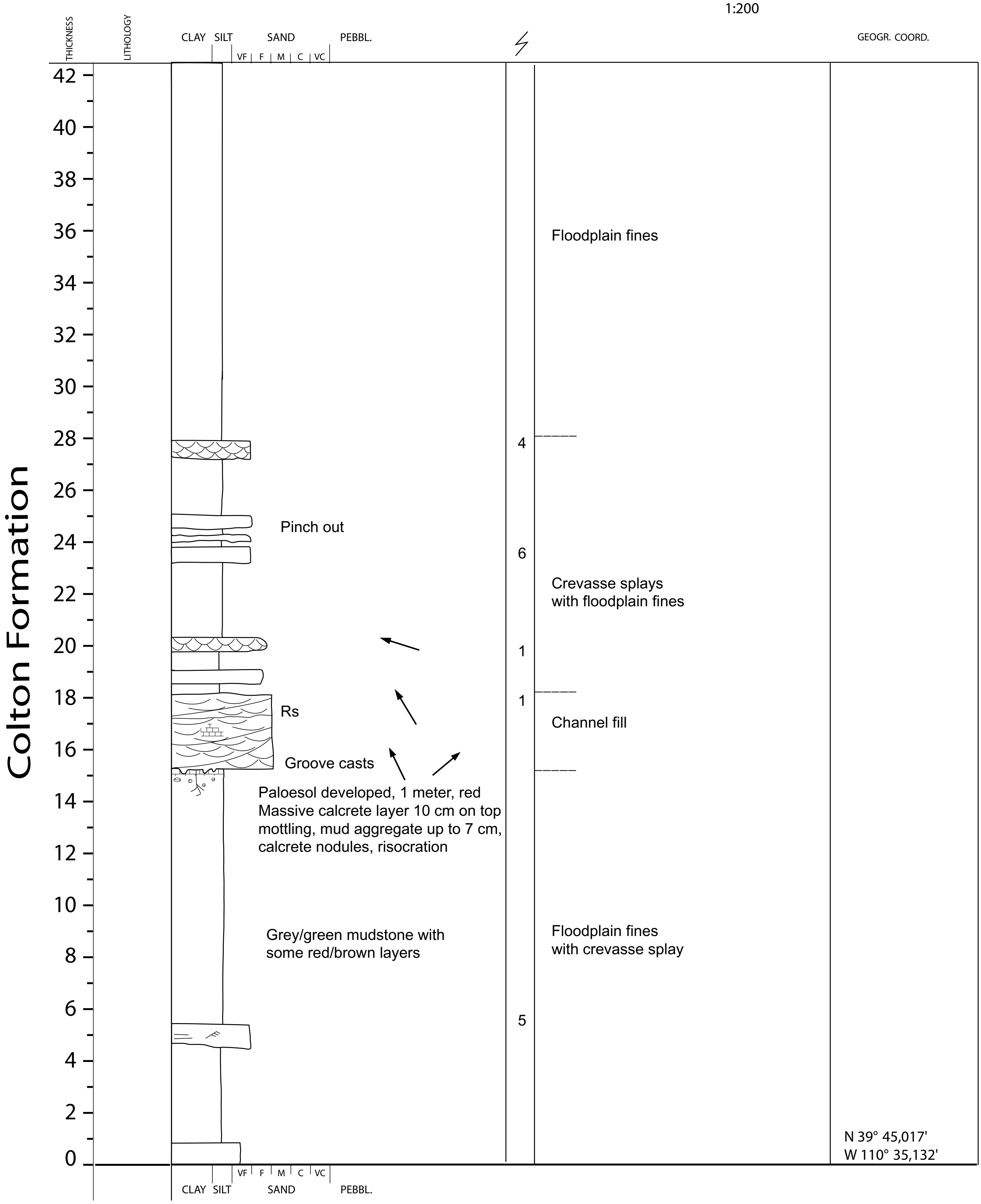
Log 17-1

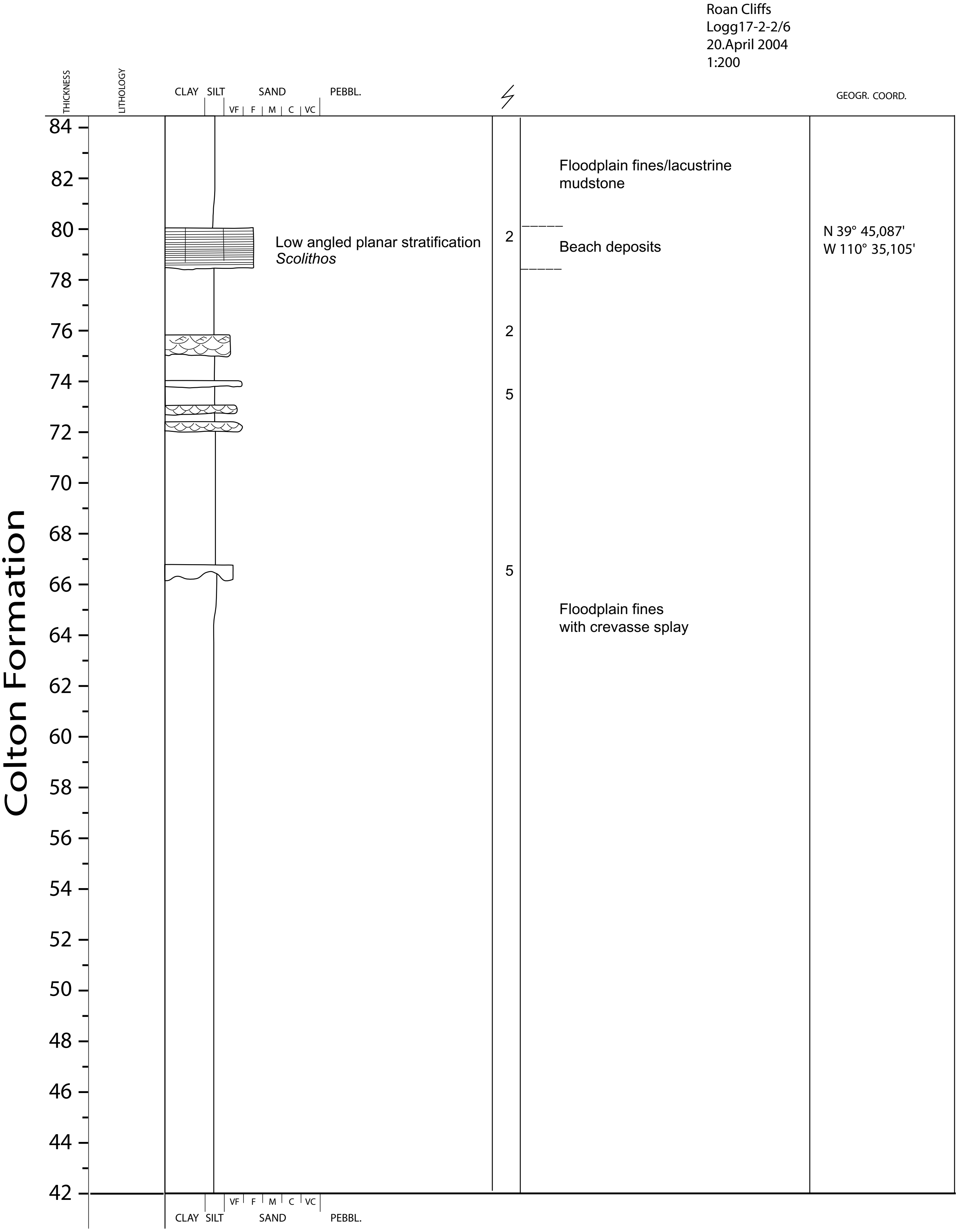
Roan Cliffs
Logg17-1-1/1
18.April 2004
1:200



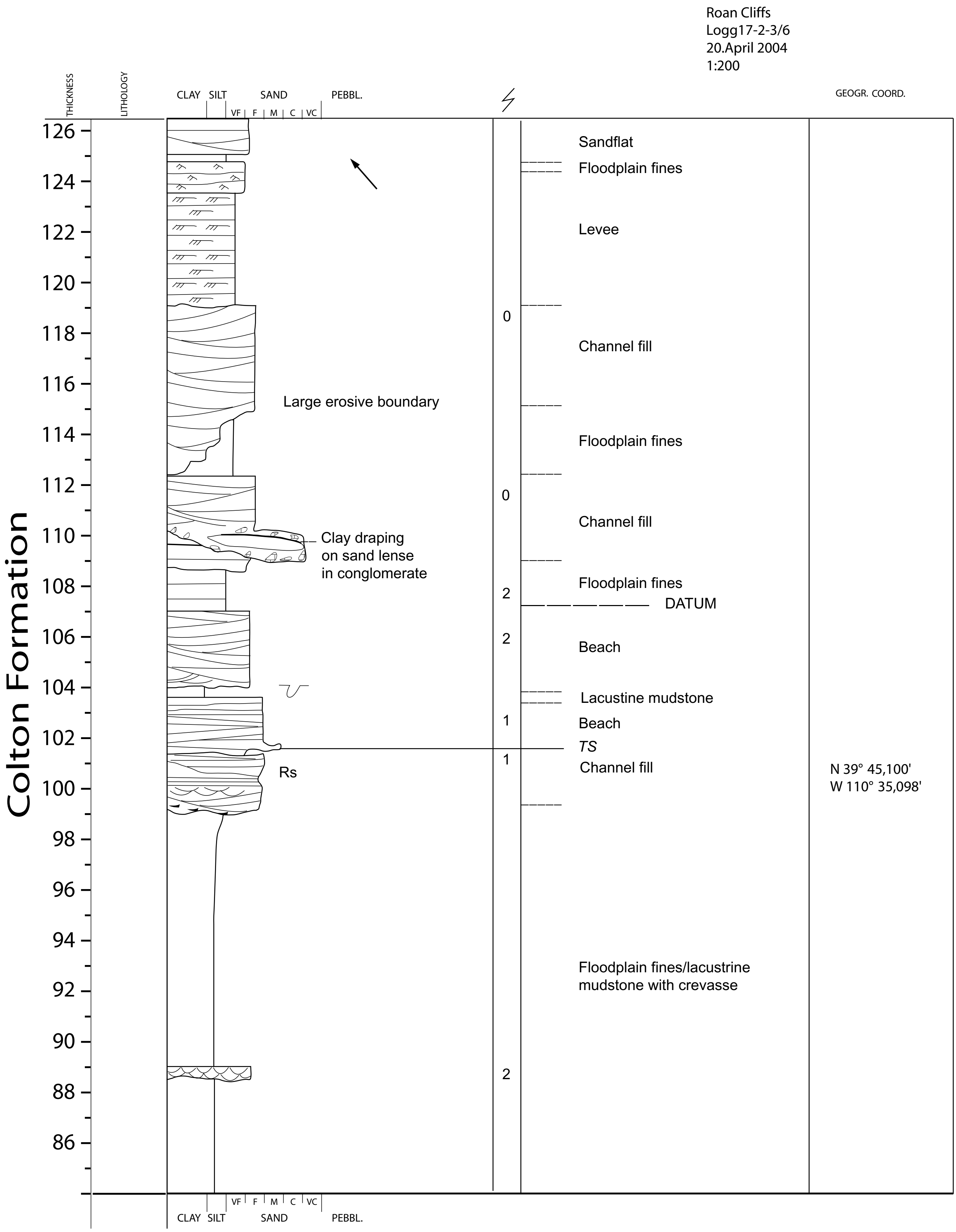
Log 17-2

Roan Cliffs
Logg17-2-1/6
20.April 2004
1:200

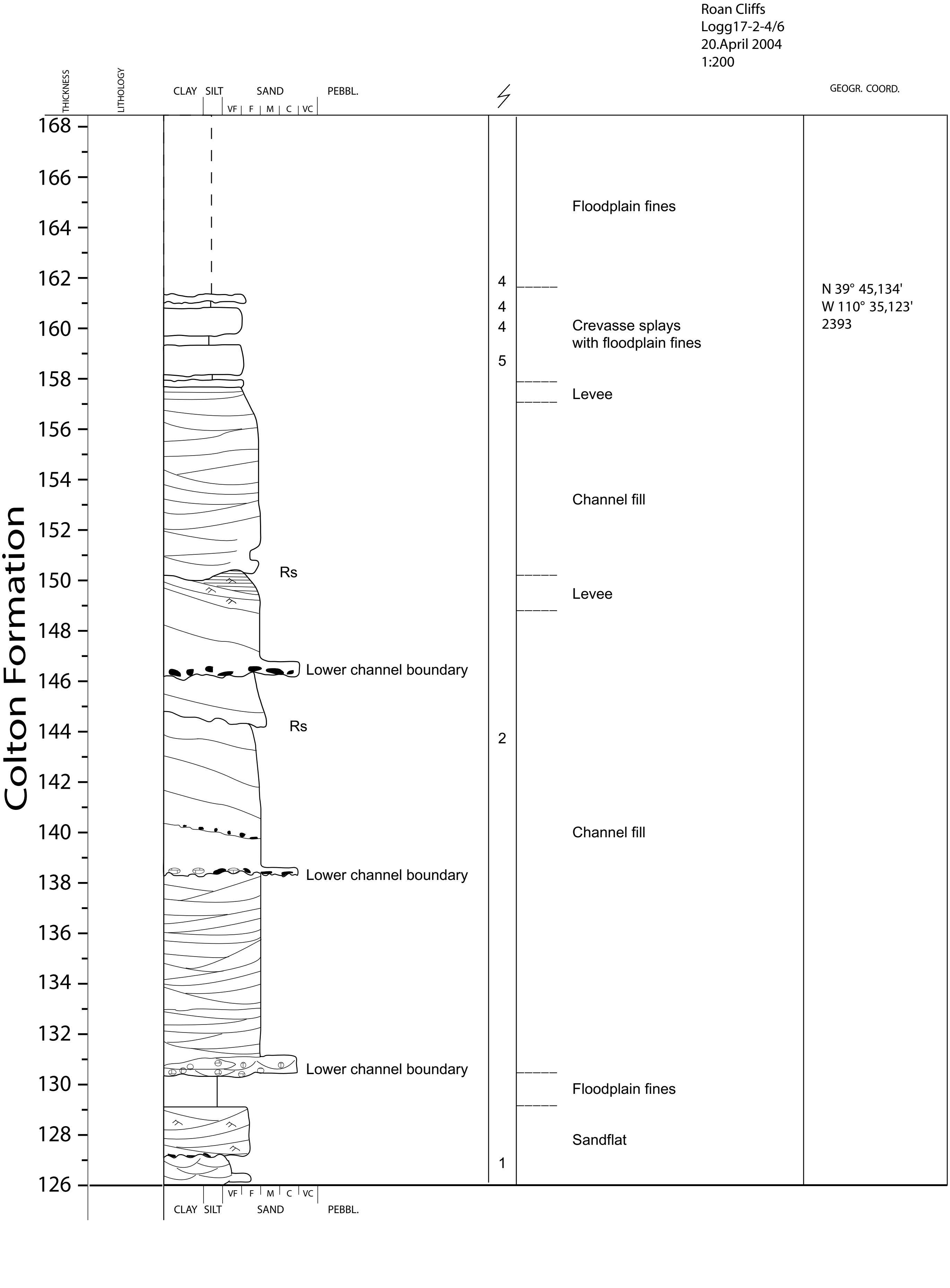




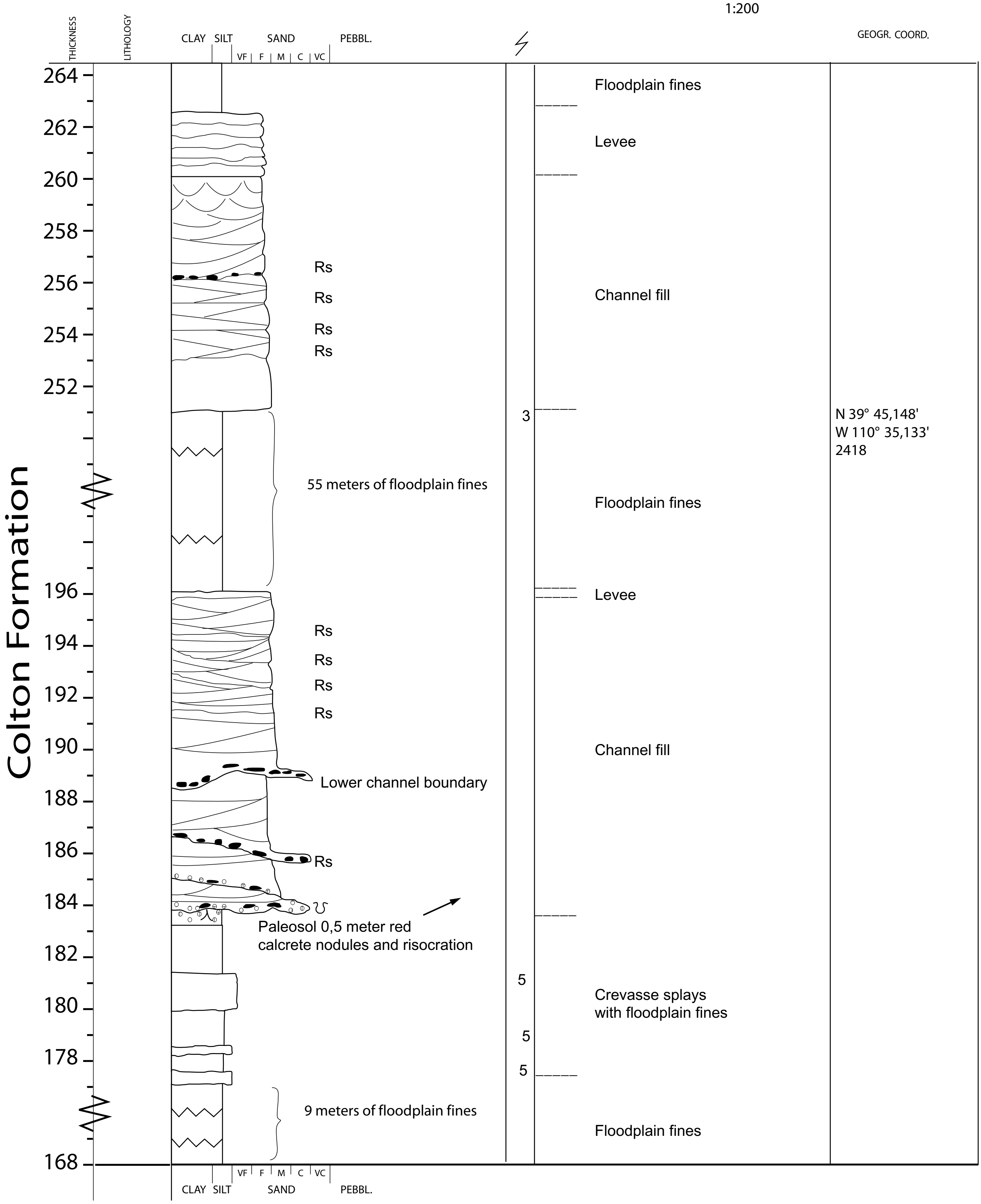
Roan Cliffs
Logg17-2-2/6
20.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
GEOGR. COORD.
Colton Formation
84
82
80
78
76
74
72
70
68
66
64
62
60
58
56
54
52
50
48
46
44
42
Low angled planar stratification
Scolithos
2
2
5
5
Floodplain fines/lacustrine mudstone
Beach deposits
Floodplain fines with crevasse splay
N 39° 45,087'
W 110° 35,105'
VF F M C VC
CLAY SILT SAND PEBBL.



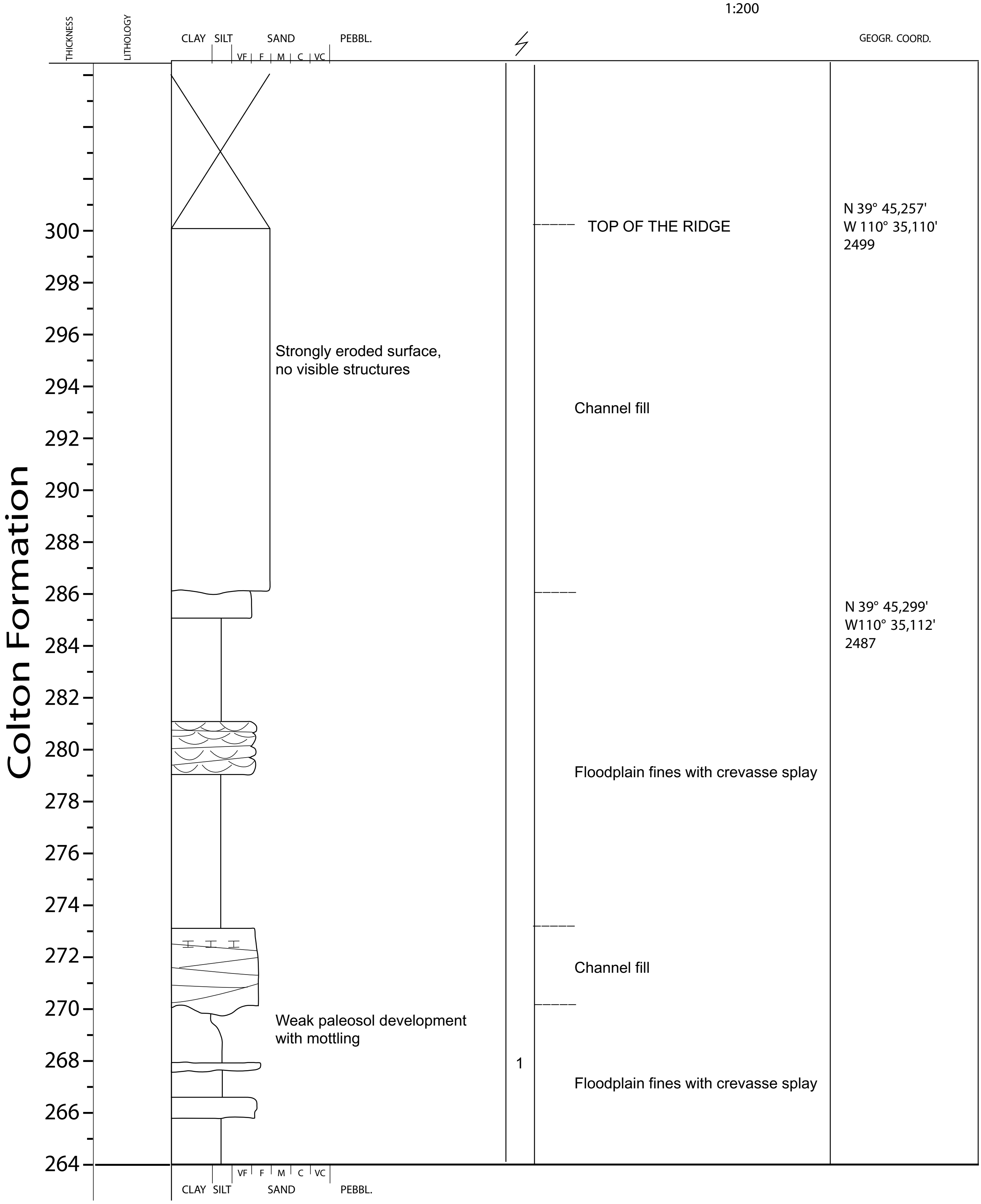
Roan Cliffs
Logg17-2-3/6
20.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
GEOGR. COORD.
Colton Formation
126 124 122 120 118 116 114 112 110 108 106 104 102 100 98 96 94 92 90 88 86
Large erosive boundary
Clay draping on sand lense in conglomerate
Rs
0 0 2 2 1 1 2
Sandflat
Floodplain fines
Levee
Channel fill
Floodplain fines
Channel fill
Floodplain fines
DATUM
Beach
Lacustrine mudstone
Beach
TS
Channel fill
Floodplain fines/lacustrine mudstone with crevasse
N 39° 45,100'
W 110° 35,098'


Roan Cliffs
Logg17-2-4/6
20.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
GEOGR. COORD.
Colton Formation
168
166
164
162
160
158
156
154
152
150
148
146
144
142
140
138
136
134
132
130
128
126
Floodplain fines
4
4
4
5
Crevasse splays
with floodplain fines
N 39° 45,134'
W 110° 35,123'
2393
Levee
Channel fill
Rs
Levee
Lower channel boundary
Rs
2
Channel fill
Lower channel boundary
Lower channel boundary
Floodplain fines
Sandflat
1
CLAY SILT SAND PEBBL.
VF F M C VC

Roan Cliffs
Logg17-2-5/6
20.April 2004
1:200

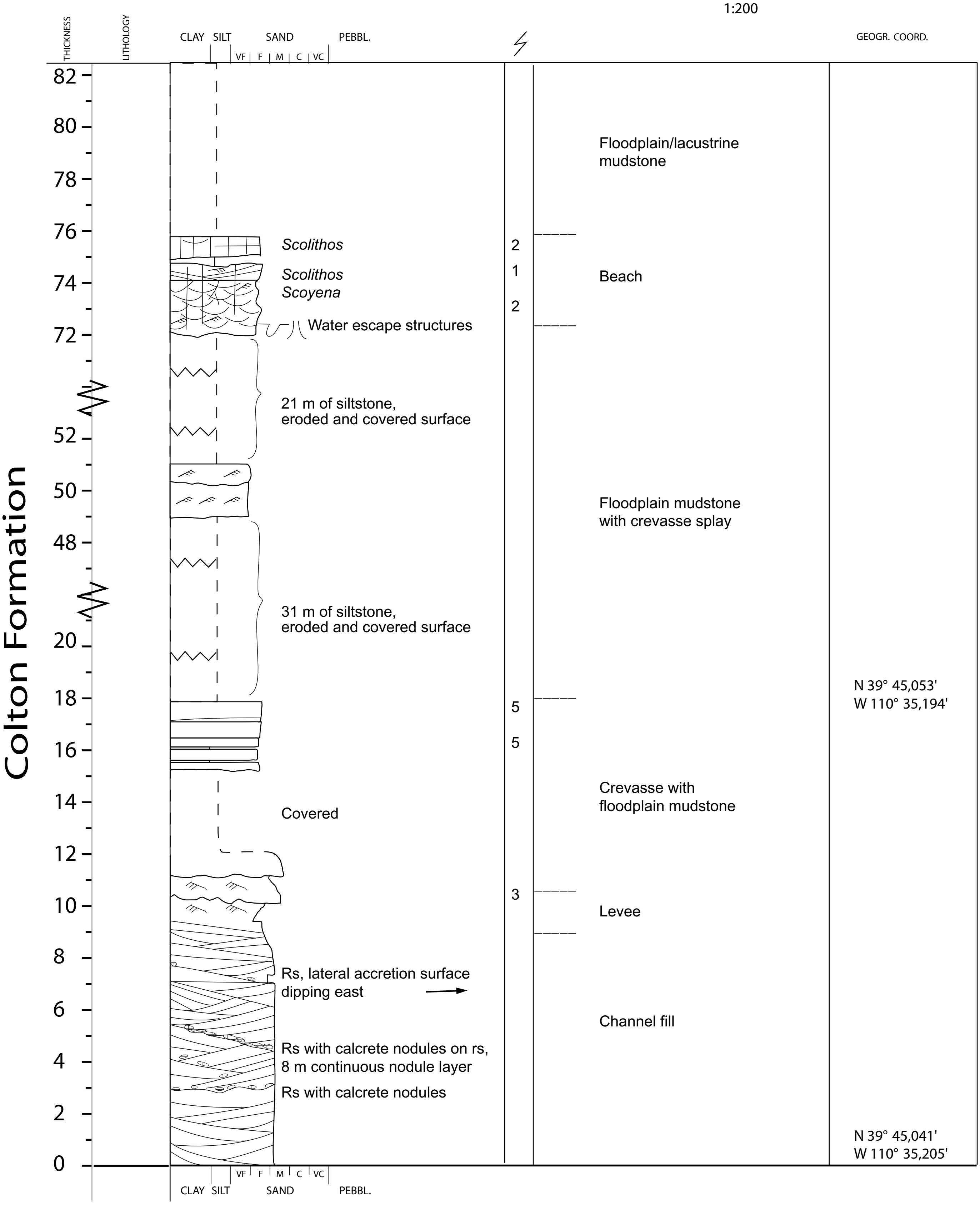


Roan Cliffs
Logg17-2-6/6
24.April 2004
1:200

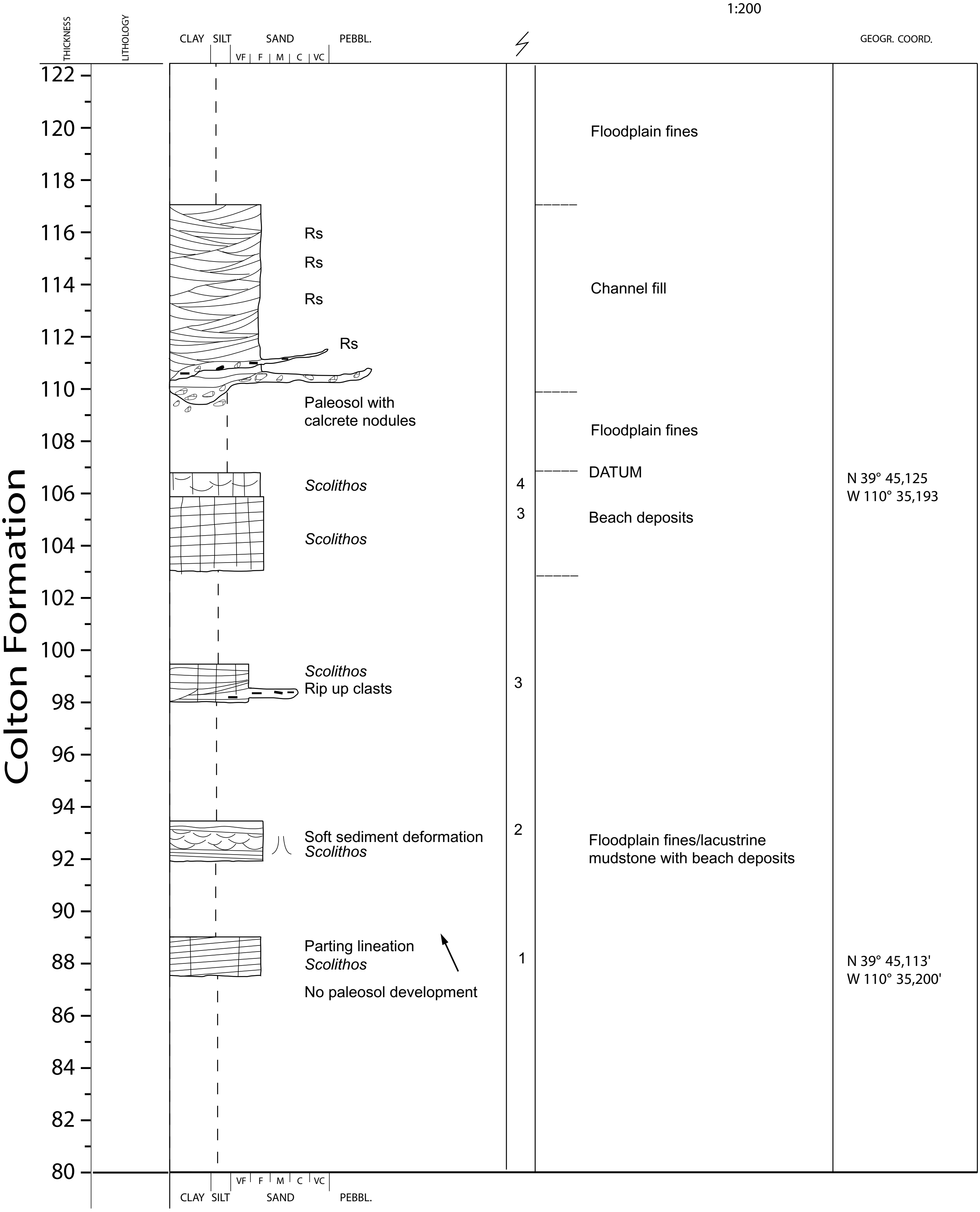


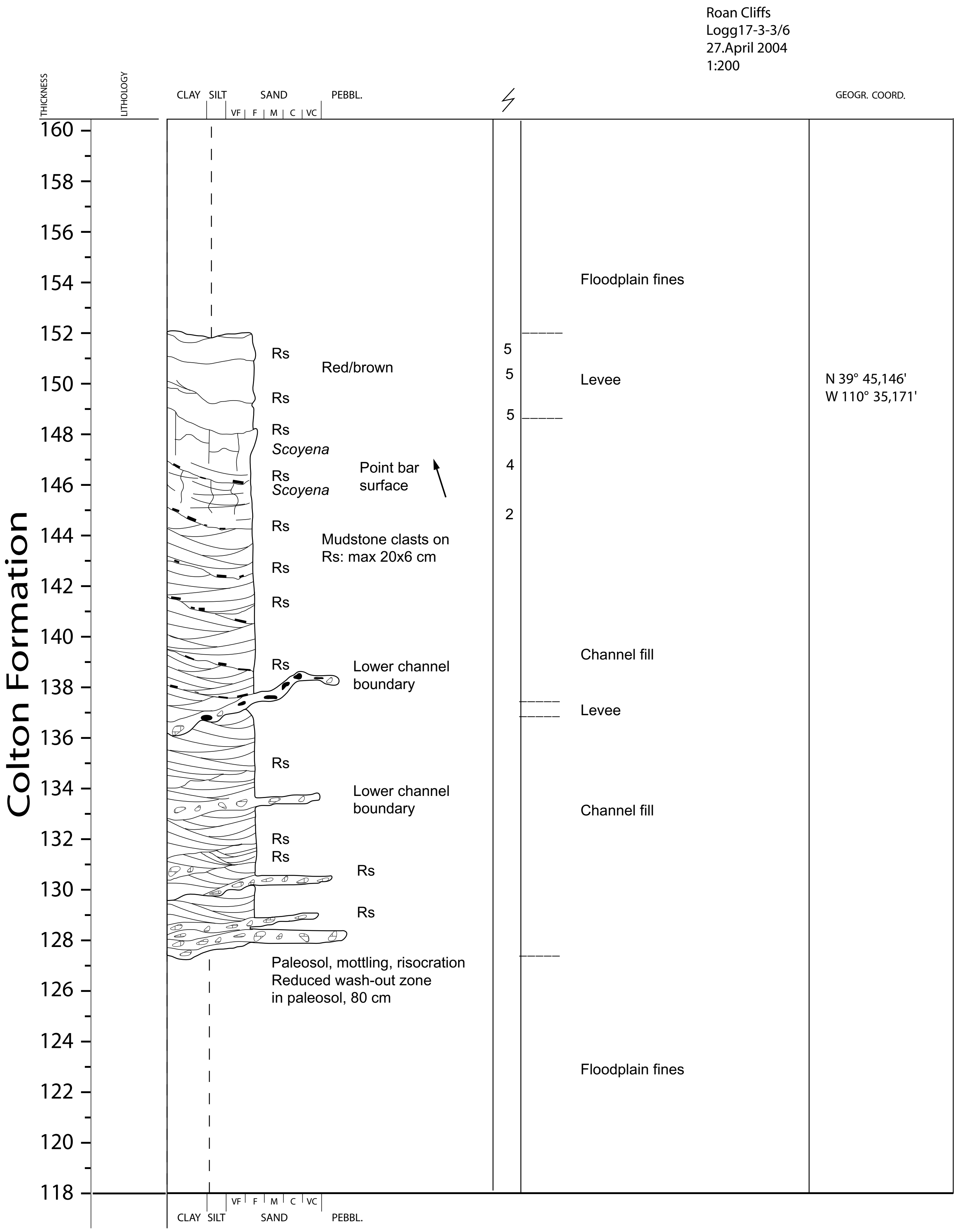
Log 17-3

Roan Cliffs
Logg17-3-1/6
21.April 2004
1:200

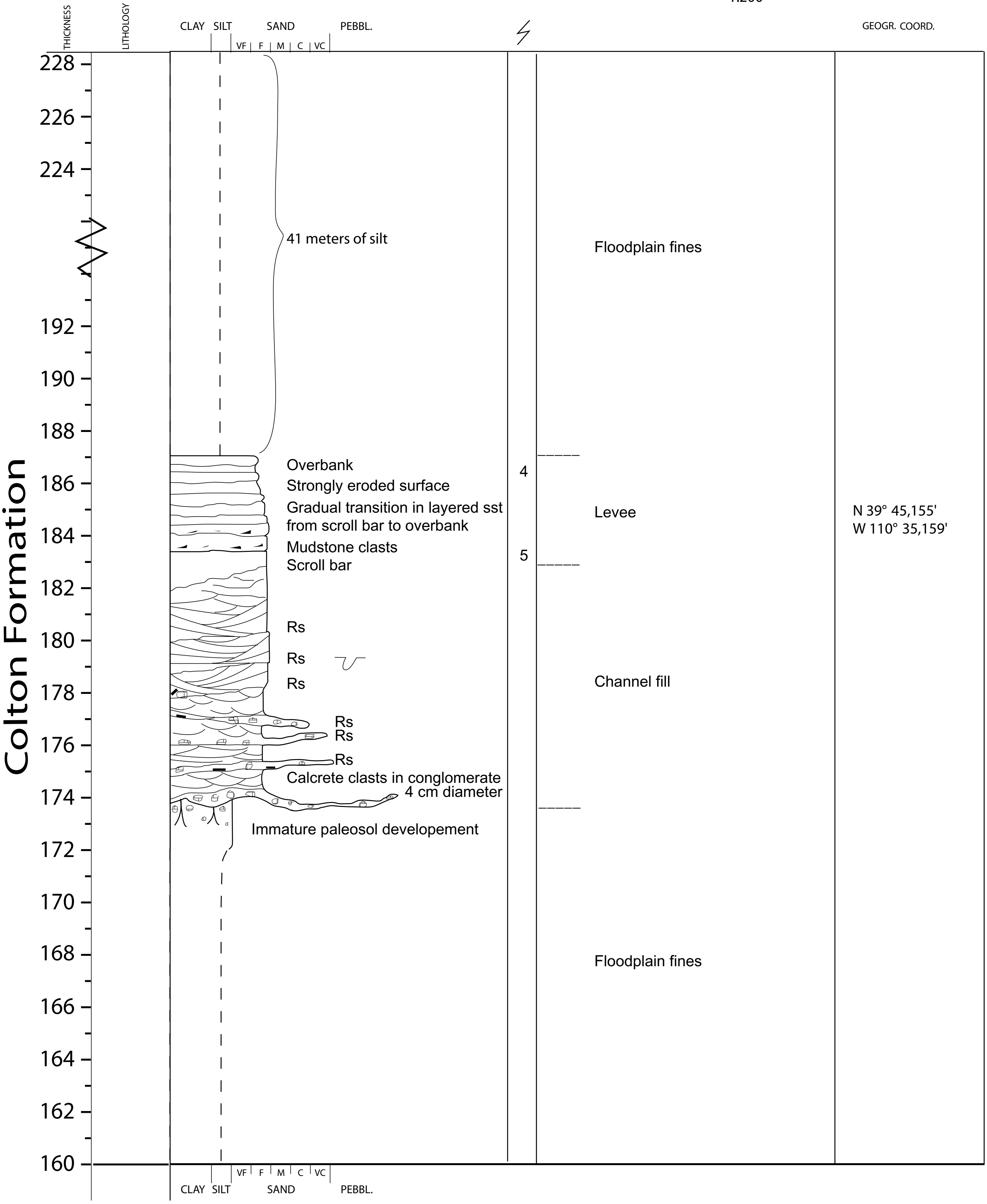


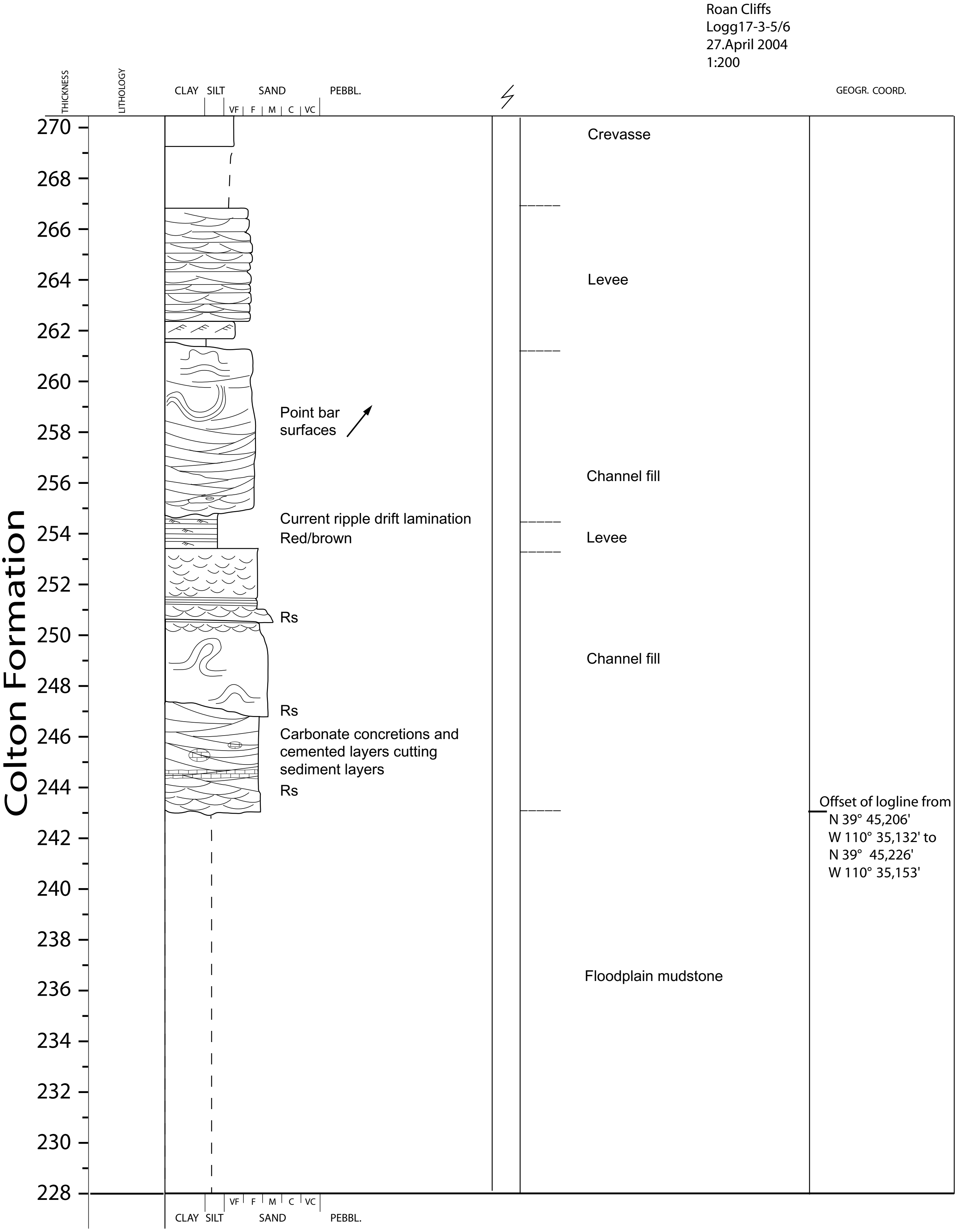
Roan Cliffs
Logg17-3-2/6
27.April 2004
1:200




Roan Cliffs
Logg17-3-3/6
27.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
GEOGR. COORD.
Colton Formation
160
158
156
154
152
150
148
146
144
142
140
138
136
134
132
130
128
126
124
122
120
118
Floodplain fines
Rs
Red/brown
5
5
Levee
N 39° 45,146'
W 110° 35,171'
Rs
5
Rs
Scoyena
Point bar
surface
4
Rs
Scoyena
2
Rs
Mudstone clasts on
Rs: max 20x6 cm
Rs
Rs
Channel fill
Rs
Lower channel
boundary
Levee
Rs
Lower channel
boundary
Channel fill
Rs
Rs
Rs
Rs
Paleosol, mottling, risocration
Reduced wash-out zone
in paleosol, 80 cm
Floodplain fines
CLAY SILT SAND PEBBL.

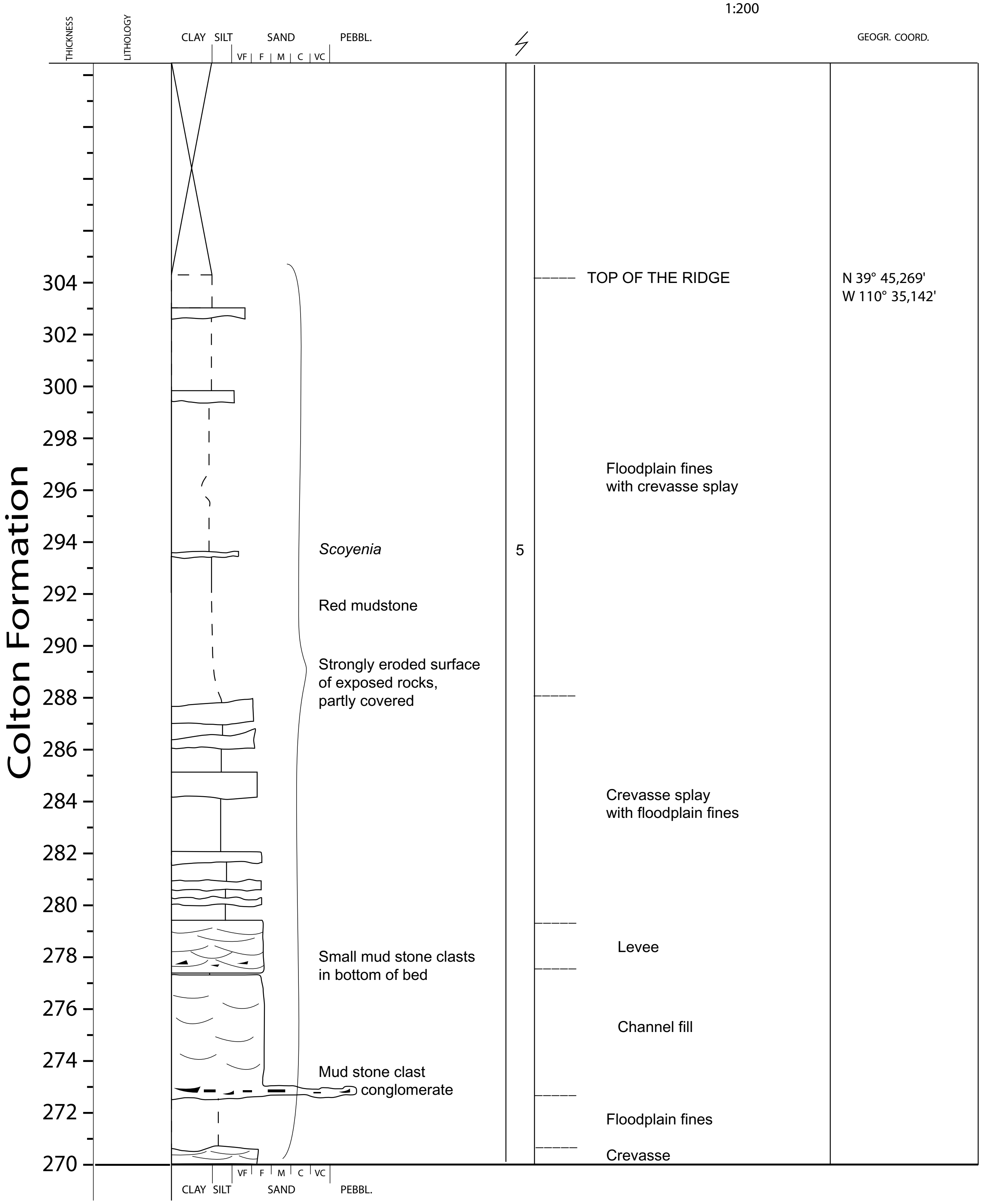
Roan Cliffs
Logg17-3-4/6
27.April 2004
1:200





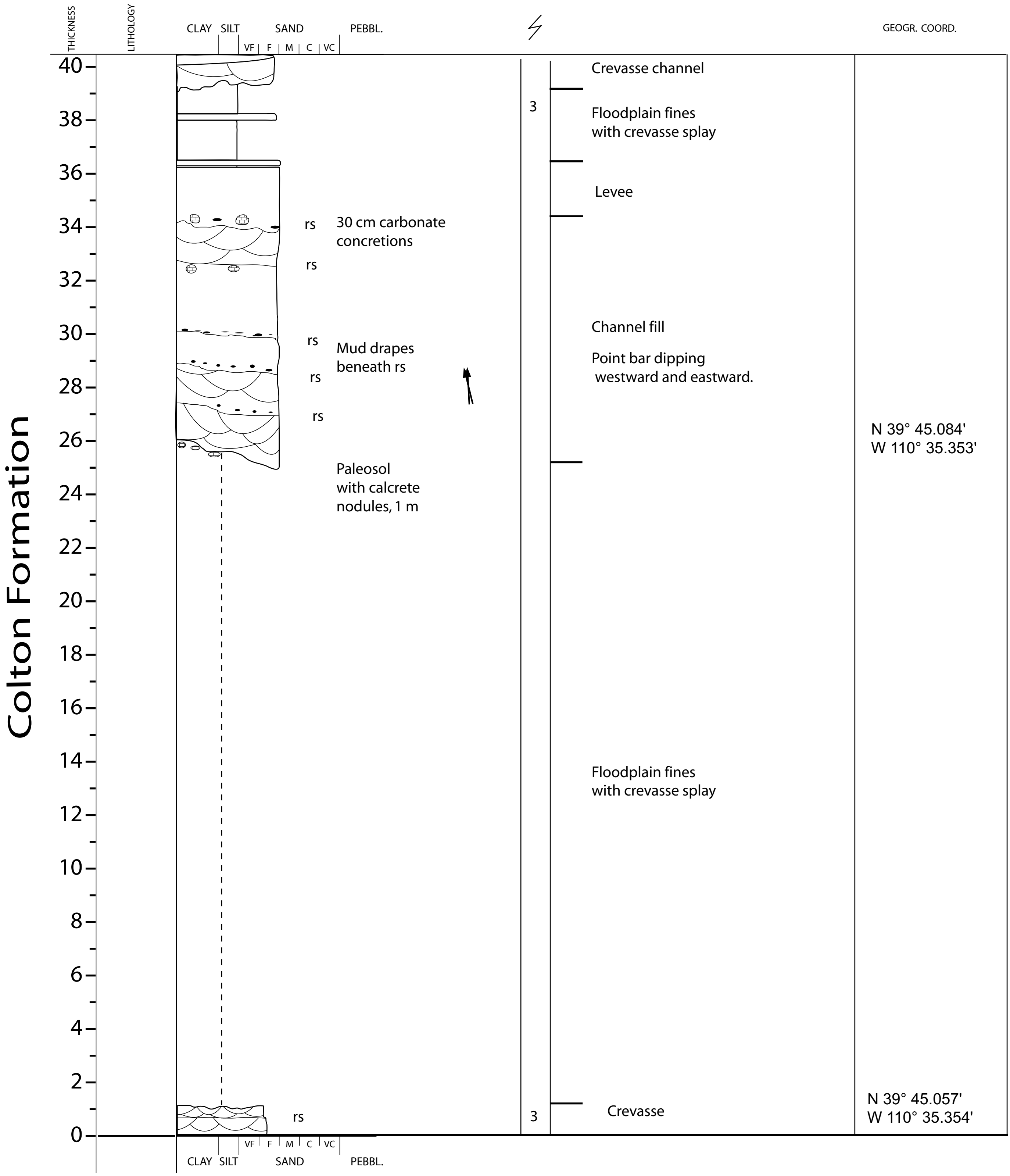
Roan Cliffs
Logg17-3-5/6
27.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY
SILT
SAND
PEBBL.
VF
F
M
C
VC
GEOGR. COORD.
Colton Formation
270
268
266
264
262
260
258
256
254
252
250
248
246
244
242
240
238
236
234
232
230
228
Crevasse
Levee
Point bar
surfaces
Channel fill
Current ripple drift lamination
Red/brown
Levee
Rs
Channel fill
Rs
Carbonate concretions and
cemented layers cutting
sediment layers
Rs
Offset of logline from
N 39° 45,206'
W 110° 35,132' to
N 39° 45,226'
W 110° 35,153'
Floodplain mudstone

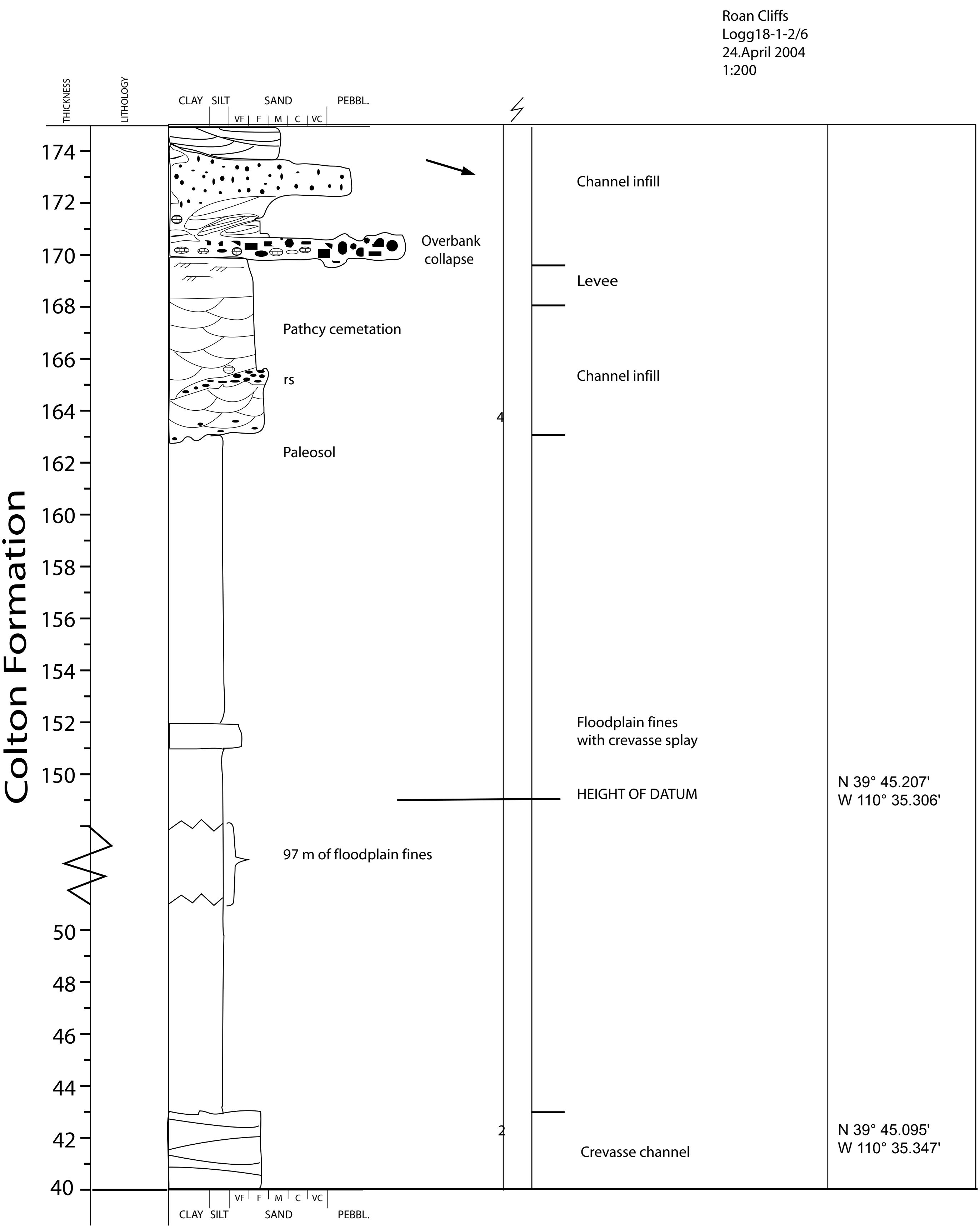
Roan Cliffs
Logg17-3-6/6
27.April 2004
1:200

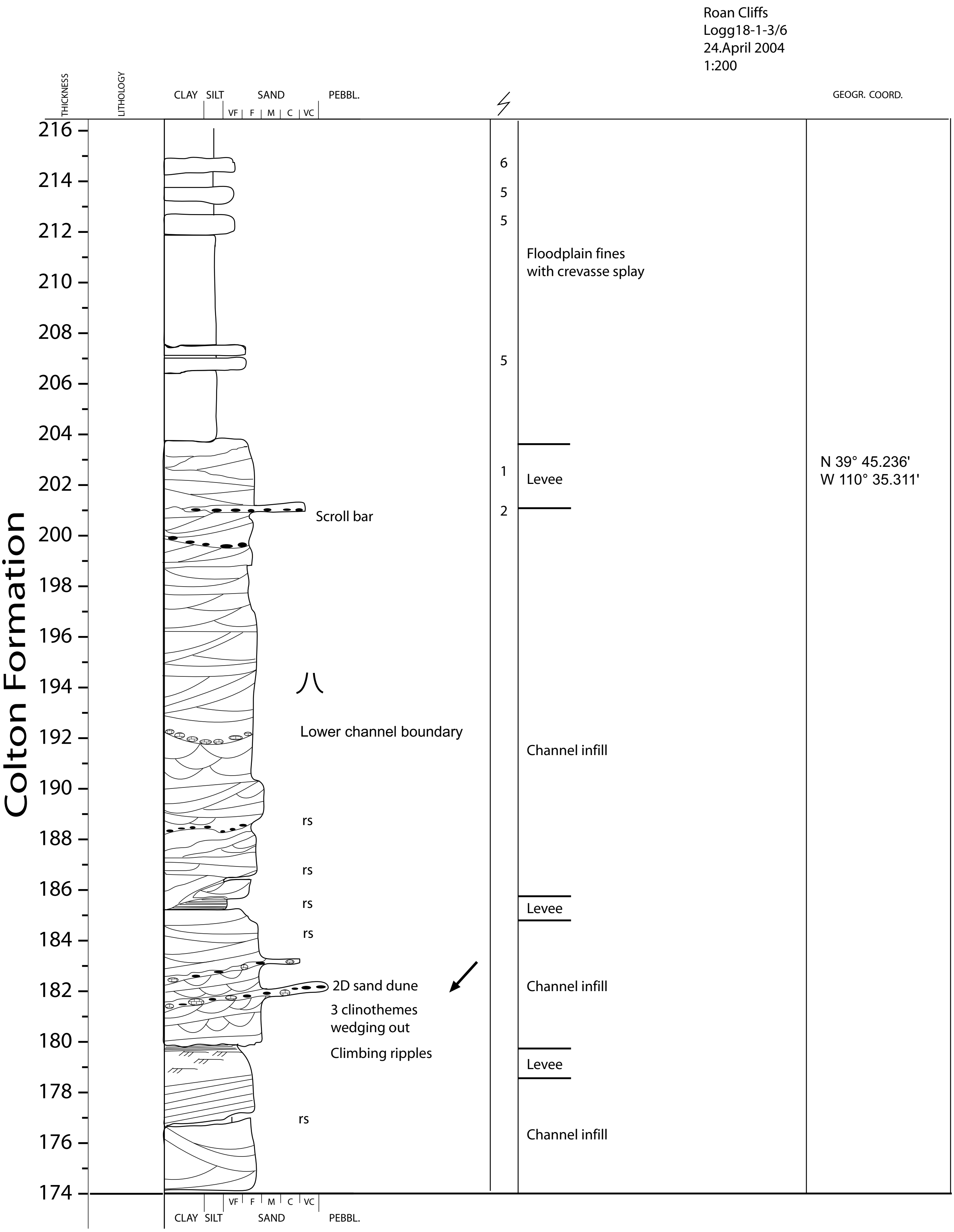


Log 18-1

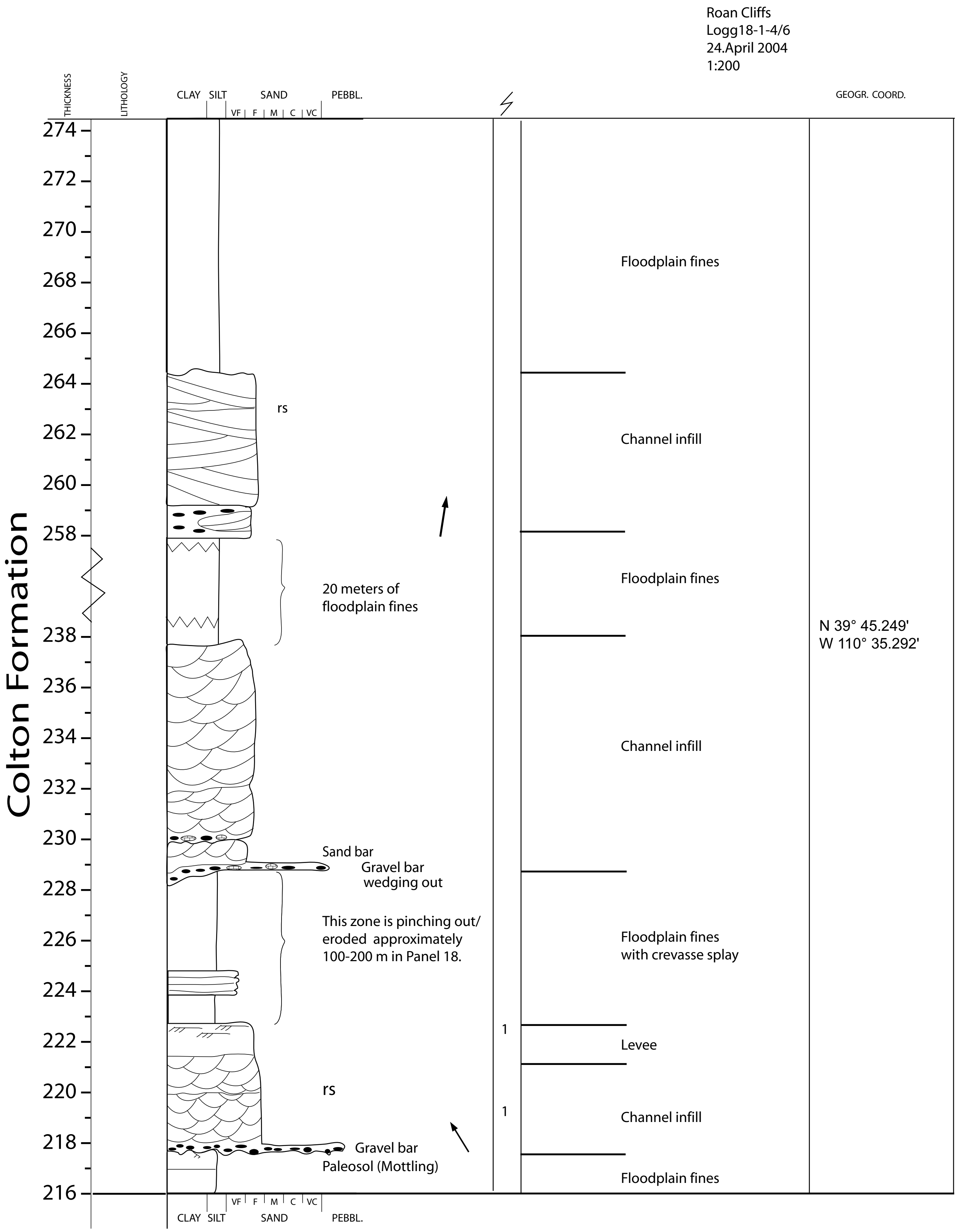
Roan Cliffs
Logg18-1-1/6
23.April 2004
1:200

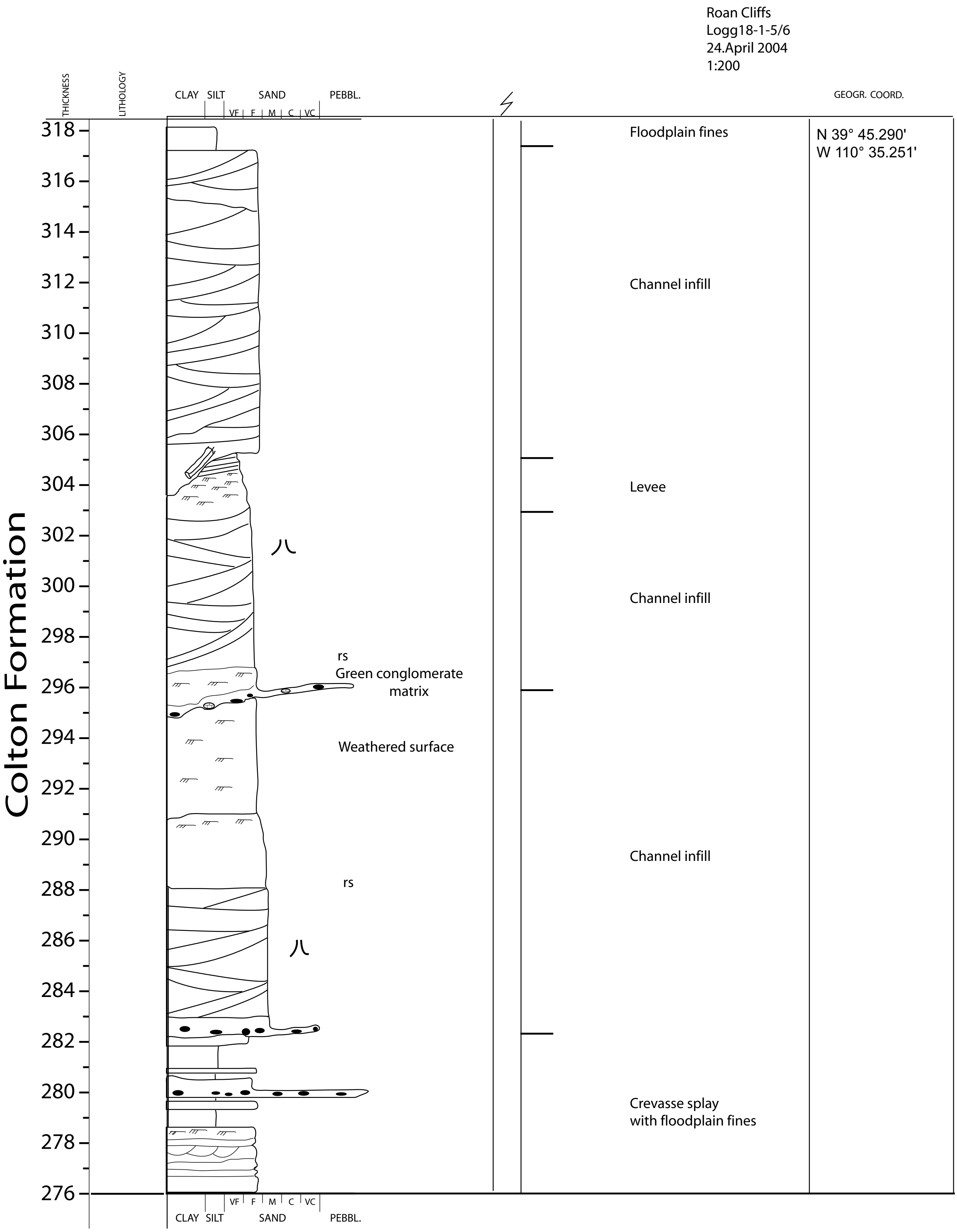



Roan Cliffs
Logg18-1-2/6
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY
SILT
SAND
PEBBL.
VF
F
M
C
VC
Colton Formation
174
172
170
168
166
164
162
160
158
156
154
152
150
50
48
46
44
42
40
Overbank
collapse
Pathcy cemetation
rs
Paleosol
97 m of floodplain fines
4
2
Channel infill
Levee
Channel infill
Floodplain fines
with crevasse splay
HEIGHT OF DATUM
Crevasse channel
N 39° 45.207'
W 110° 35.306'
N 39° 45.095'
W 110° 35.347'

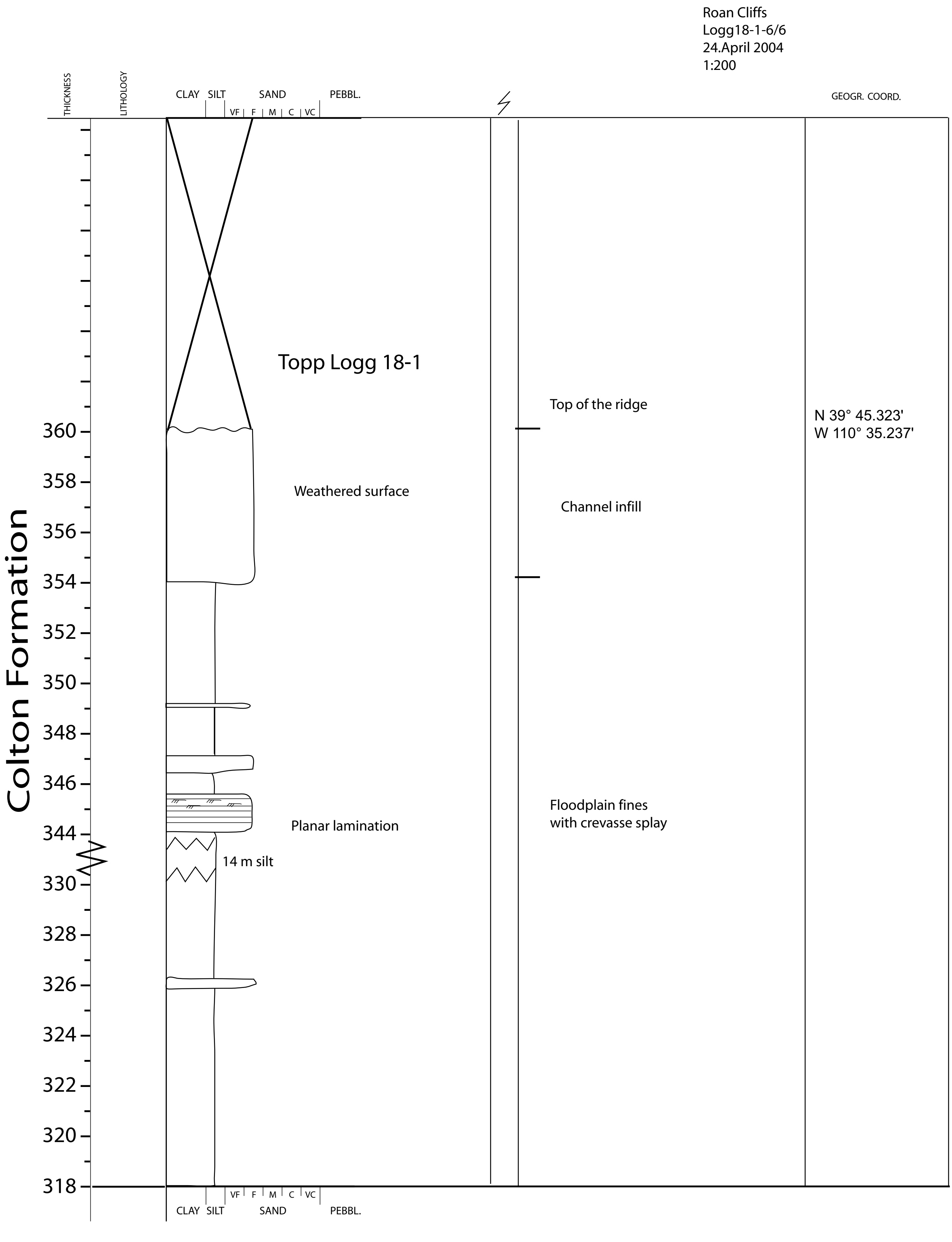


Roan Cliffs
Logg18-1-3/6
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
GEOGR. COORD.
Colton Formation
216
214
212
210
208
206
204
202
200
198
196
194
192
190
188
186
184
182
180
178
176
174
6
5
5
5
1
2
Floodplain fines
with crevasse splay
Levee
N 39° 45.236'
W 110° 35.311'
Scroll bar
Lower channel boundary
Channel infill
rs
rs
rs
Levee
rs
2D sand dune
Channel infill
3 clinothemes
wedging out
Climbing ripples
Levee
rs
Channel infill
VF F M C VC
CLAY SILT SAND PEBBL.


Roan Cliffs
Logg18-1-4/6
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY
SILT
SAND
VF
F
M
C
VC
PEBBL.
GEOGR. COORD.
Colton Formation
274
272
270
268
266
264
262
260
258
238
236
234
232
230
228
226
224
222
220
218
216
rs
20 meters of floodplain fines
Sand bar
Gravel bar wedging out
This zone is pinching out/ eroded approximately 100-200 m in Panel 18.
rs
Gravel bar
Paleosol (Mottling)
1
1
Floodplain fines
Channel infill
Floodplain fines
Channel infill
Floodplain fines with crevasse splay
Levee
Channel infill
Floodplain fines
N 39° 45.249'
W 110° 35.292'



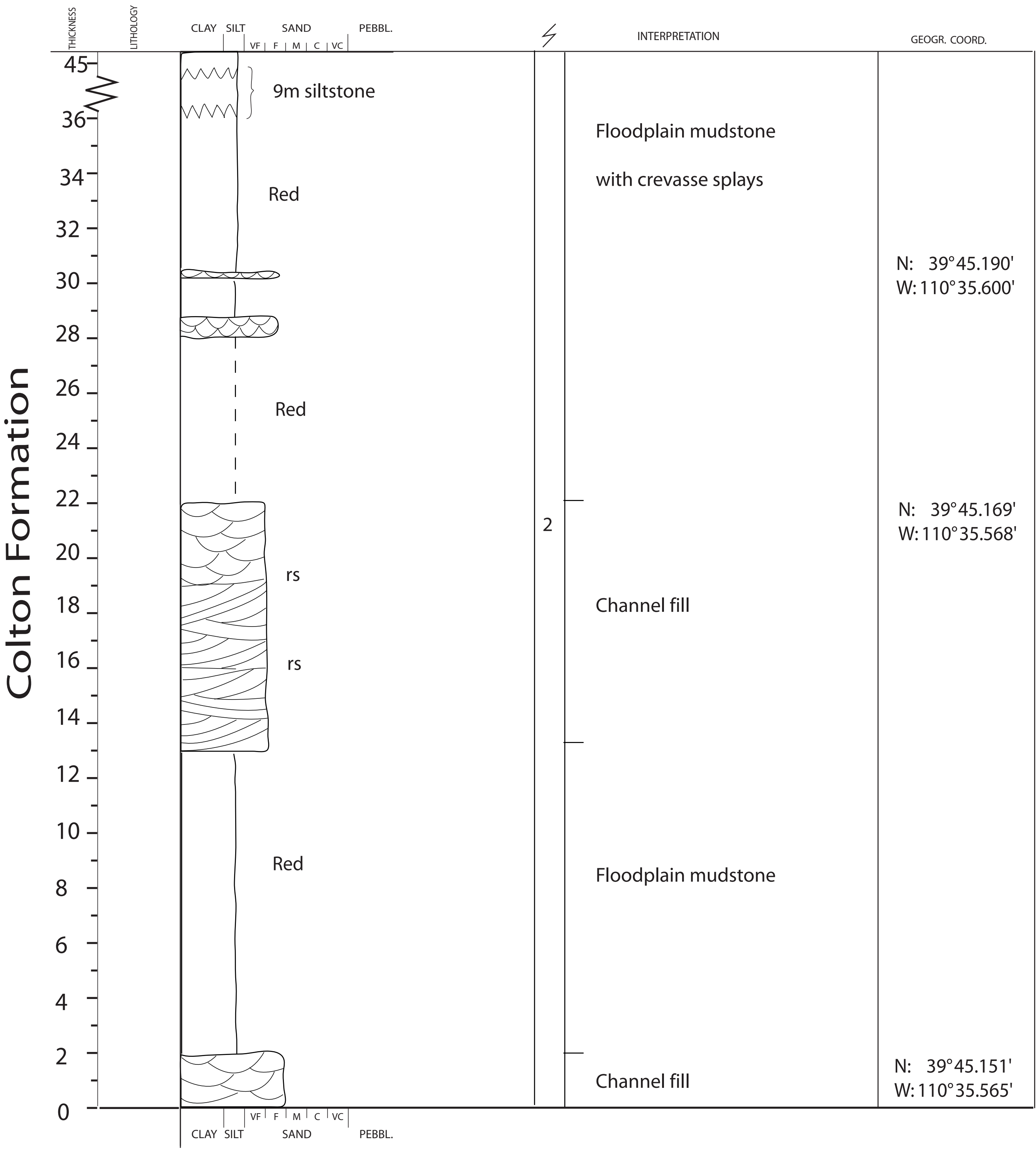
Roan Cliffs
Logg18-1-5/6
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
GEOGR. COORD.
N 39° 45.290'
W 110° 35.251'
Colton Formation
318
316
314
312
310
308
306
304
302
300
298
296
294
292
290
288
286
284
282
280
278
276
Floodplain fines
Channel infill
Levee
Channel infill
rs
Green conglomerate matrix
Weathered surface
Channel infill
rs
Crevasse splay with floodplain fines

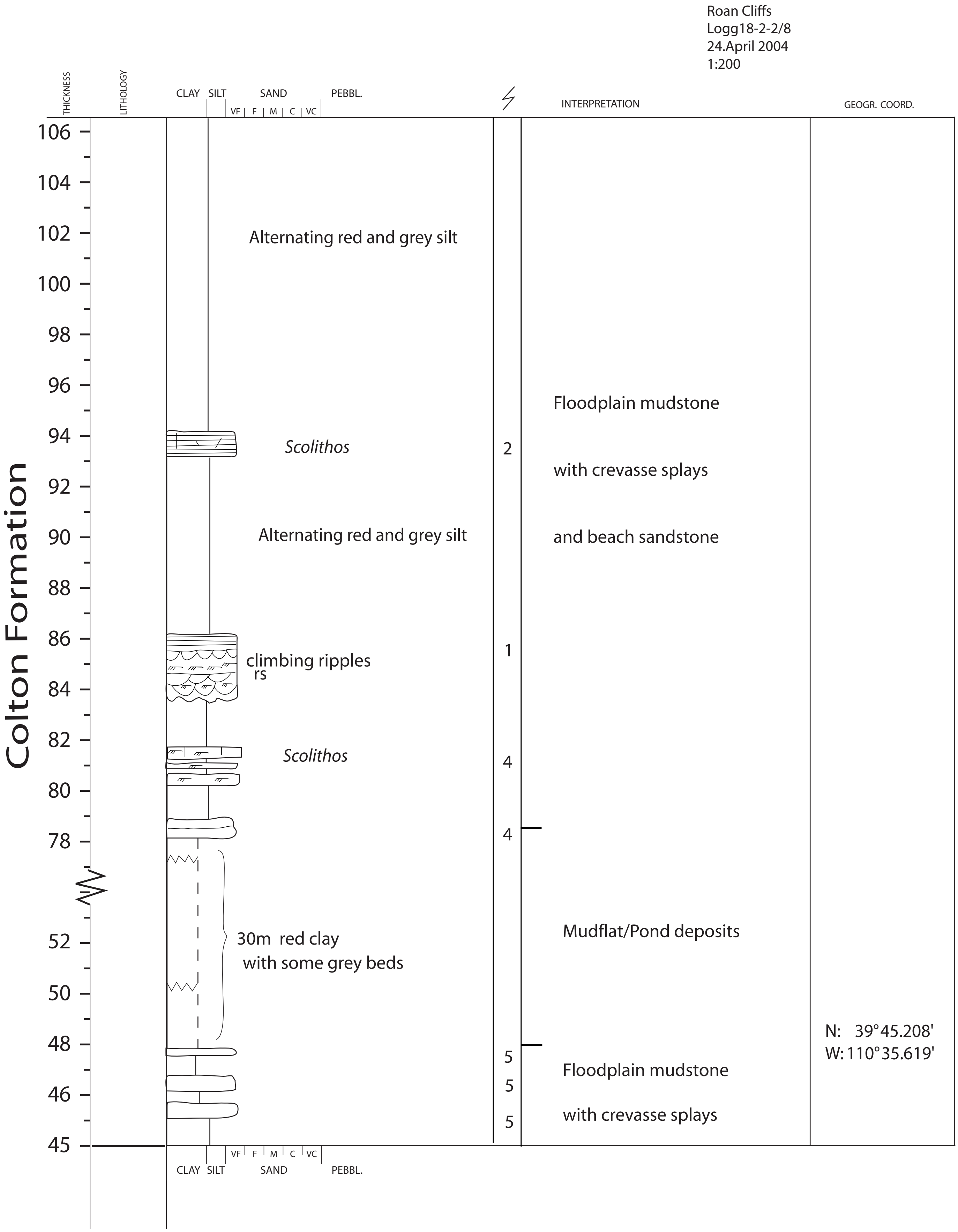


Roan Cliffs
Logg18-1-6/6
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY
SILT
SAND
PEBBL.
VF
F
M
C
VC
GEOGR. COORD.
Topp Logg 18-1
Top of the ridge
N 39° 45.323'
W 110° 35.237'
Weathered surface
Channel infill
Colton Formation
Planar lamination
Floodplain fines
with crevasse splay
14 m silt
360
358
356
354
352
350
348
346
344
330
328
326
324
322
320
318

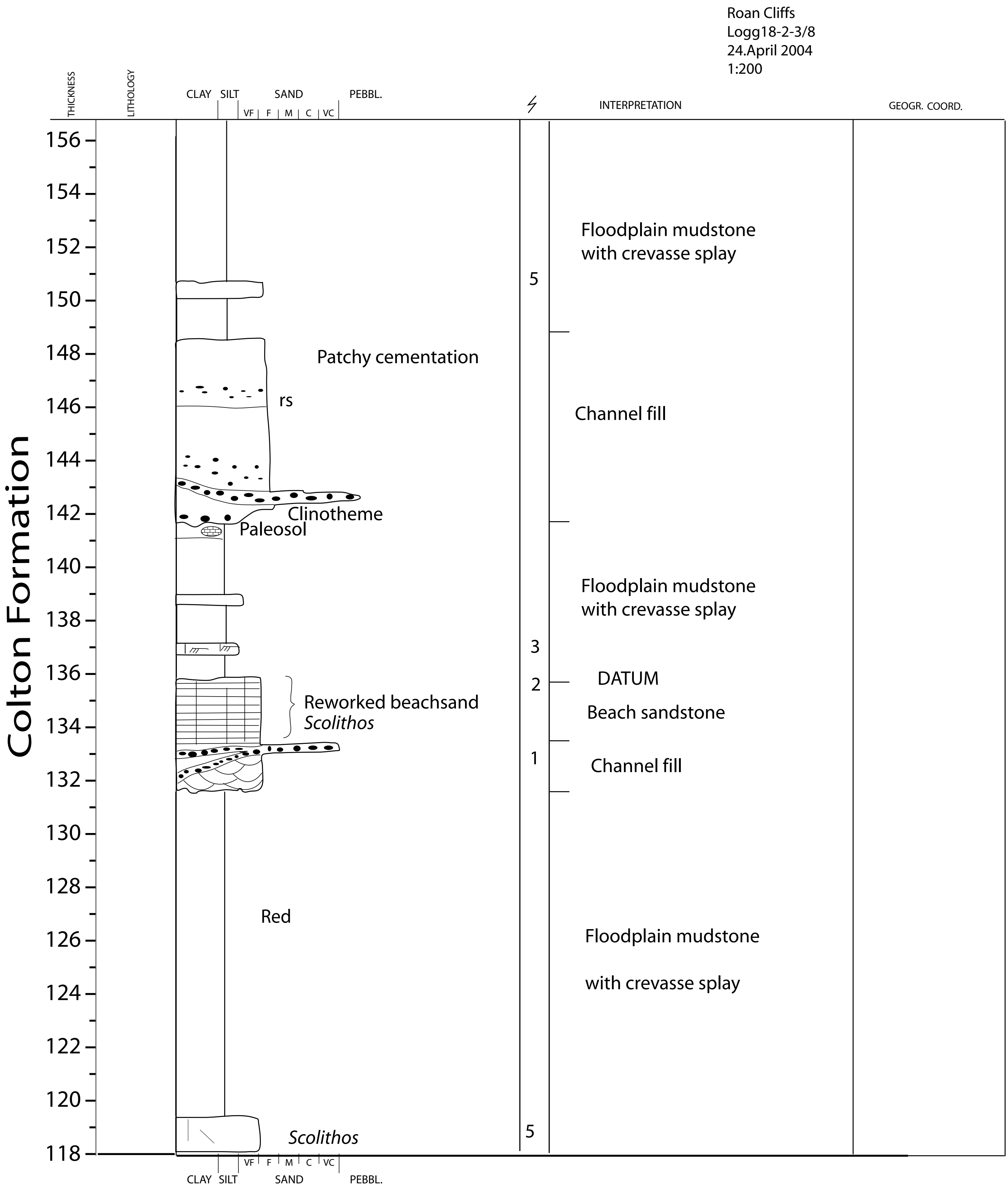
Log 18-2

Roan Cliffs
Logg18-2-1/8
24.April 2004
1:200

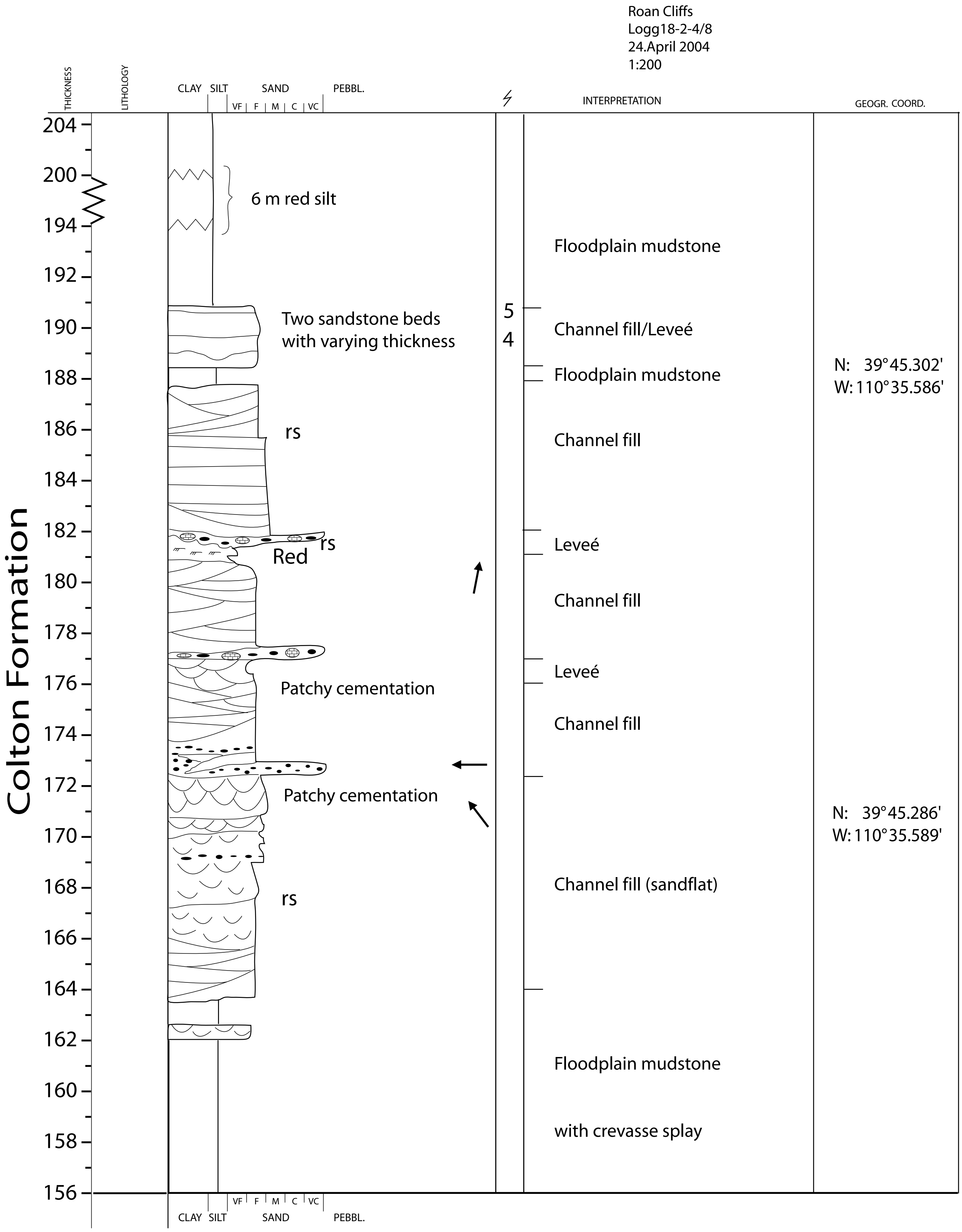




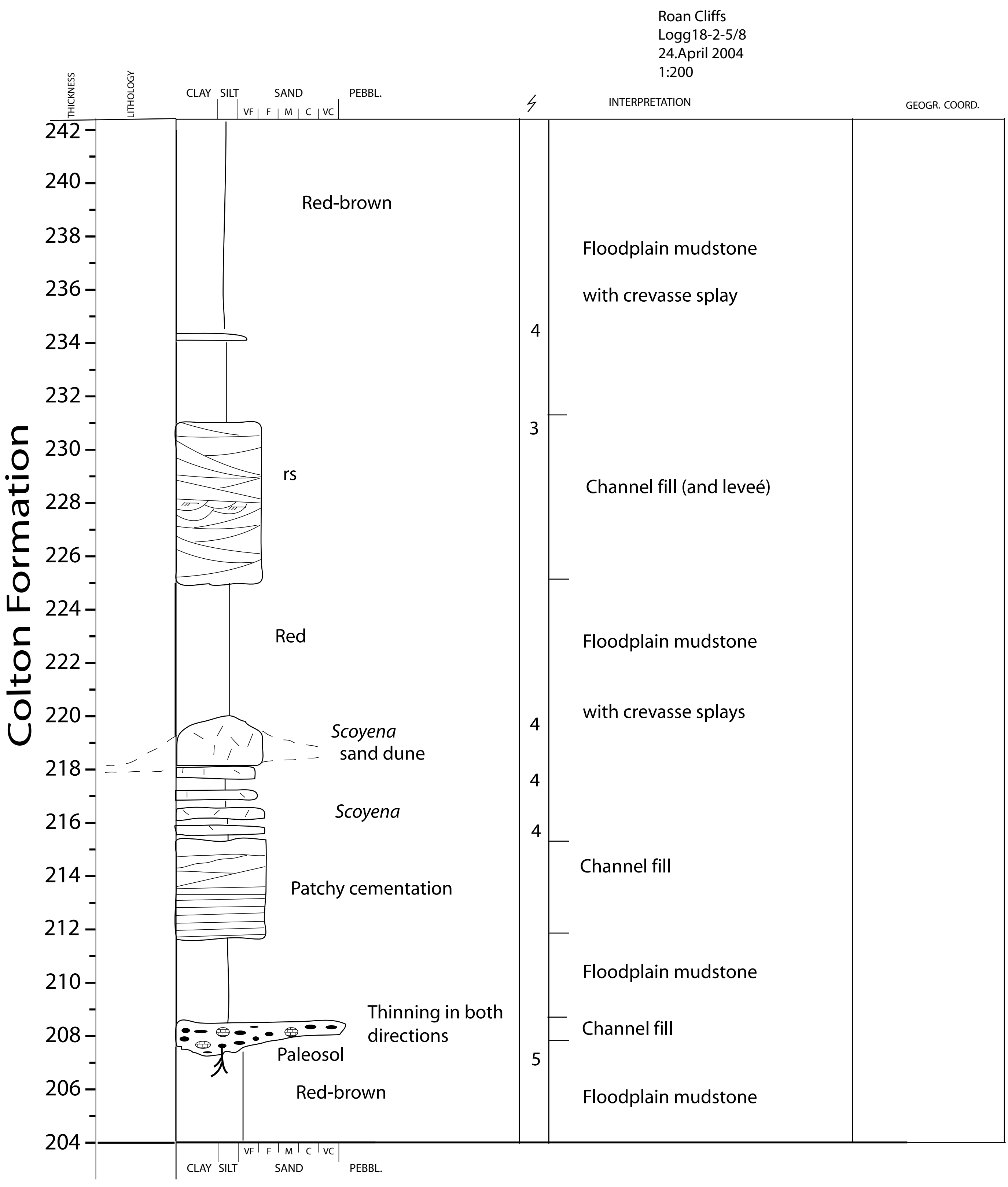
Roan Cliffs
Logg18-2-2/8
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY
SILT
SAND
VF
F
M
C
VC
PEBBL.
INTERPRETATION
GEOGR. COORD.
Colton Formation
106
104
102
100
98
96
94
92
90
88
86
84
82
80
78
52
50
48
46
45
Alternating red and grey silt
Scolithos
Alternating red and grey silt
climbing ripples
rs
Scolithos
30m red clay
with some grey beds
2
1
4
4
5
5
5
Floodplain mudstone
with crevasse splays
and beach sandstone
Mudflat/Pond deposits
Floodplain mudstone
with crevasse splays
N: 39°45.208'
W: 110°35.619'



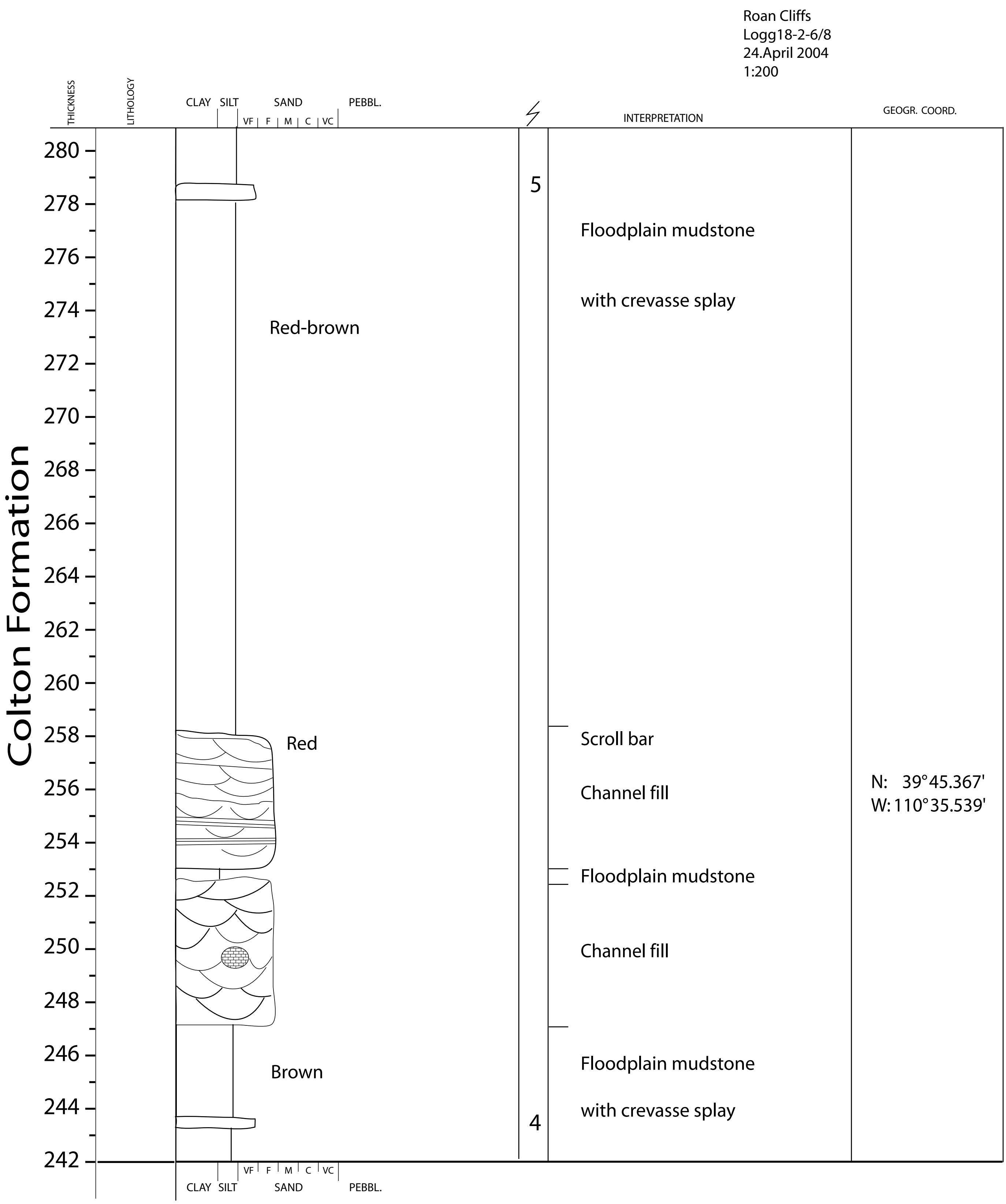
Roan Cliffs
Logg18-2-3/8
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATION
GEOGR. COORD.
Colton Formation
156
154
152
150
148
146
144
142
140
138
136
134
132
130
128
126
124
122
120
118
Patchy cementation
rs
Clinotheme
Paleosol
Reworked beachsand
Scolithos
Red
Scolithos
5
3
2
1
5
Floodplain mudstone
with crevasse splay
Channel fill
Floodplain mudstone
with crevasse splay
DATUM
Beach sandstone
Channel fill
Floodplain mudstone
with crevasse splay



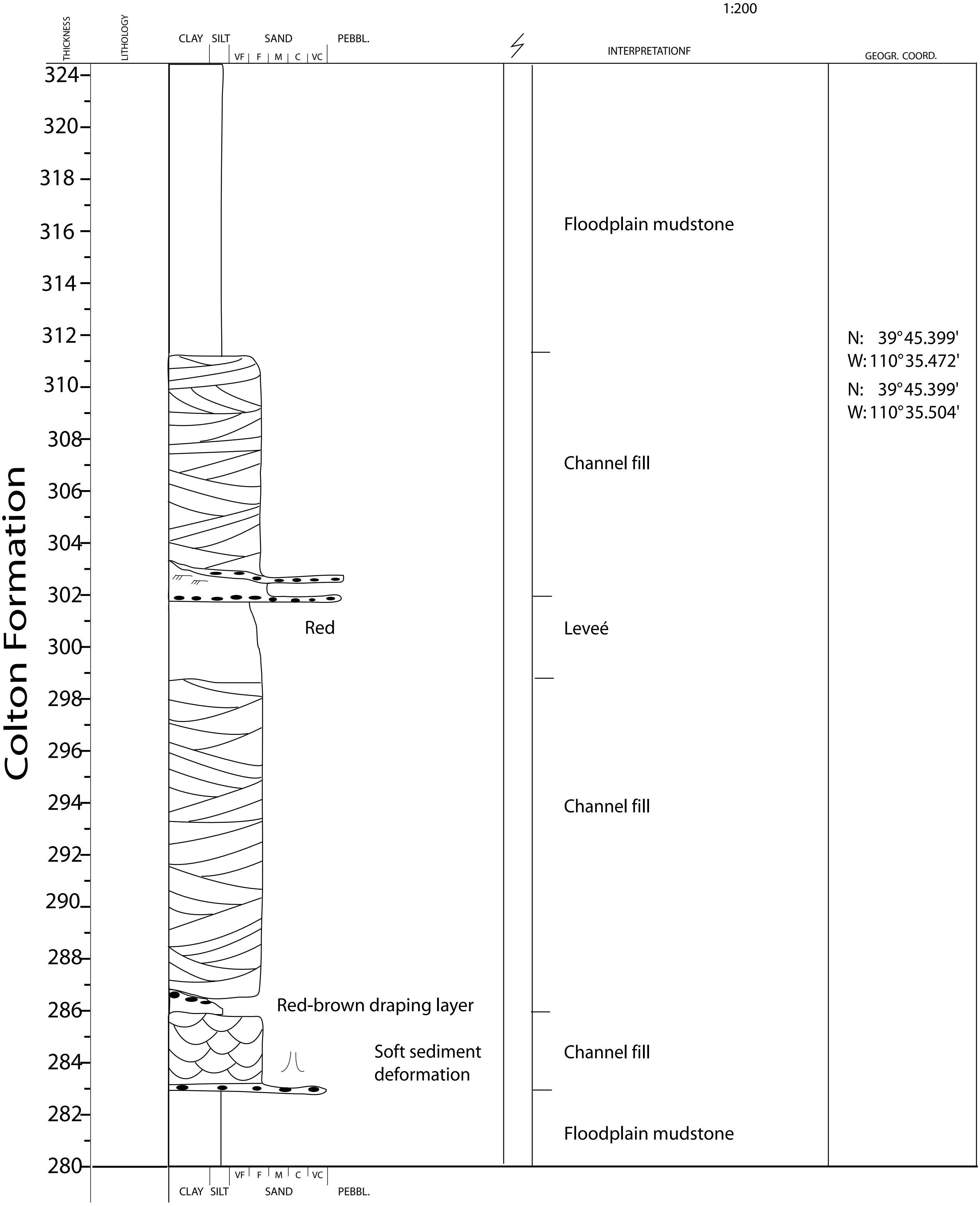
Roan Cliffs
Logg18-2-4/8
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATION
GEOGR. COORD.
Colton Formation
204
200
194
192
190
188
186
184
182
180
178
176
174
172
170
168
166
164
162
160
158
156
6 m red silt
Two sandstone beds with varying thickness
rs
rs
Red
Patchy cementation
Patchy cementation
rs
5
4
Floodplain mudstone
Channel fill/Leveé
Floodplain mudstone
Channel fill
Leveé
Channel fill
Leveé
Channel fill
Channel fill (sandflat)
Floodplain mudstone
with crevasse splay
N: 39°45.302'
W: 110°35.586'
N: 39°45.286'
W: 110°35.589'
CLAY SILT SAND PEBBL.
VF F M C VC



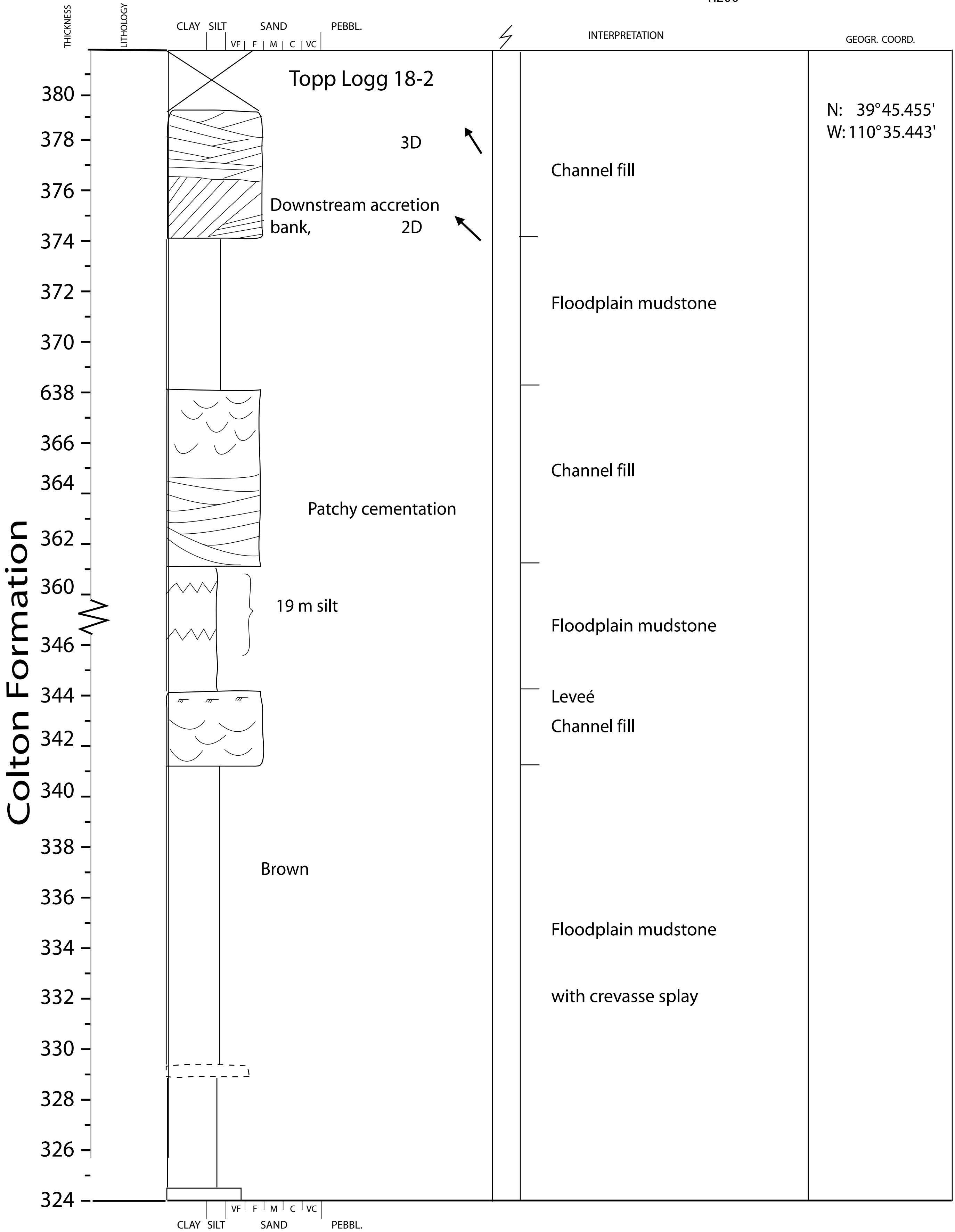
Roan Cliffs
Logg18-2-5/8
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATION
GEOGR. COORD.
Colton Formation
242
240
238
236
234
232
230
228
226
224
222
220
218
216
214
212
210
208
206
204
Red-brown
rs
Red
Scoyena
sand dune
Scoyena
Patchy cementation
Thinning in both directions
Paleosol
Red-brown
4
3
4
4
4
5
Floodplain mudstone
with crevasse splay
Channel fill (and leveé)
Floodplain mudstone
with crevasse splays
Channel fill
Floodplain mudstone
Channel fill
Floodplain mudstone
VF F M C VC
CLAY SILT SAND PEBBL.


Roan Cliffs
Logg18-2-6/8
24.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATION
GEOGR. COORD.
Colton Formation
280
278
276
274
272
270
268
266
264
262
260
258
256
254
252
250
248
246
244
242
5
Floodplain mudstone
with crevasse splay
Red-brown
Red
Scroll bar
Channel fill
N: 39°45.367'
W: 110°35.539'
Floodplain mudstone
Channel fill
Brown
Floodplain mudstone
with crevasse splay
4
VF F M C VC
CLAY SILT SAND PEBBL.

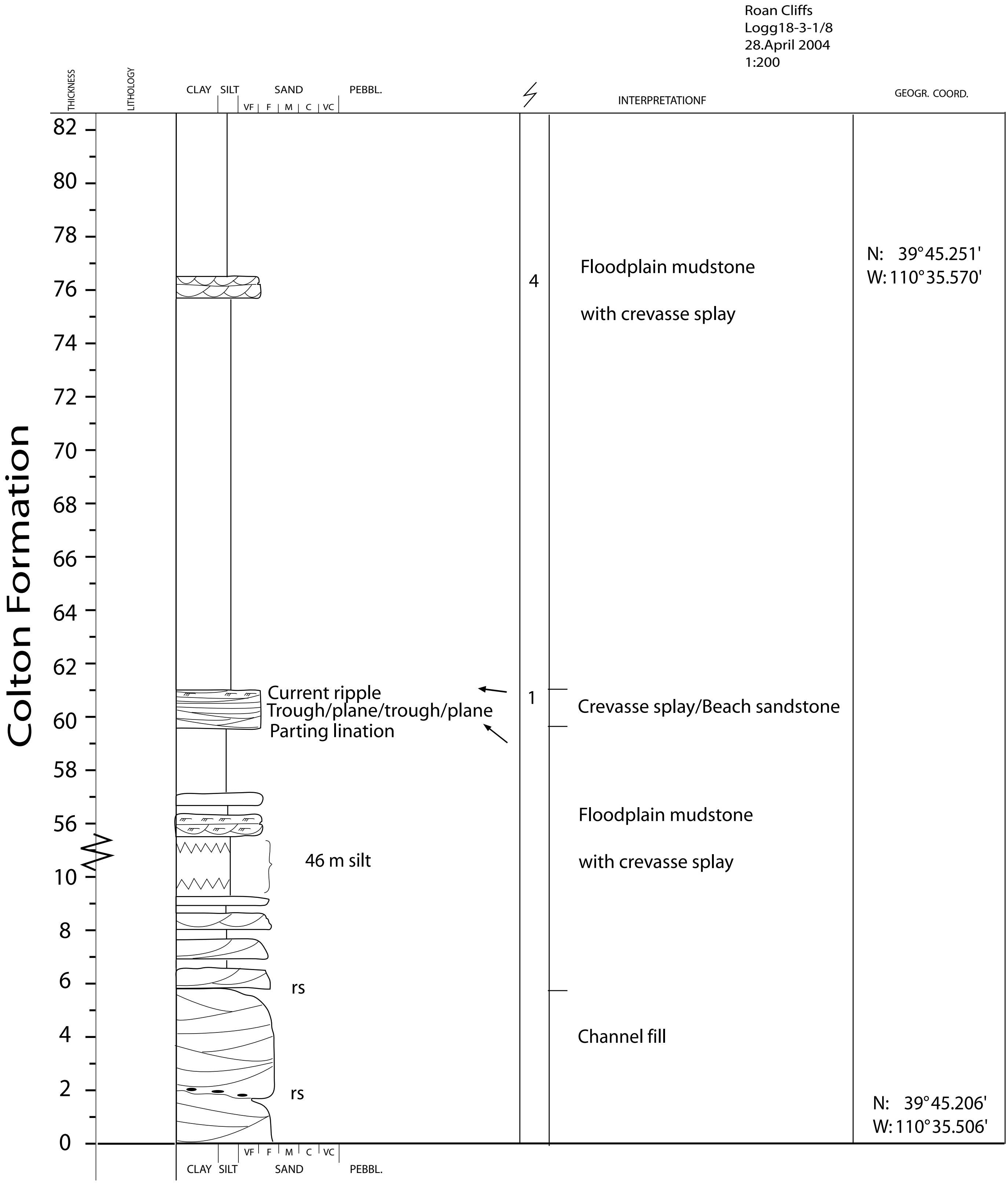
Roan Cliffs
Logg18-2-7/8
24.April 2004
1:200



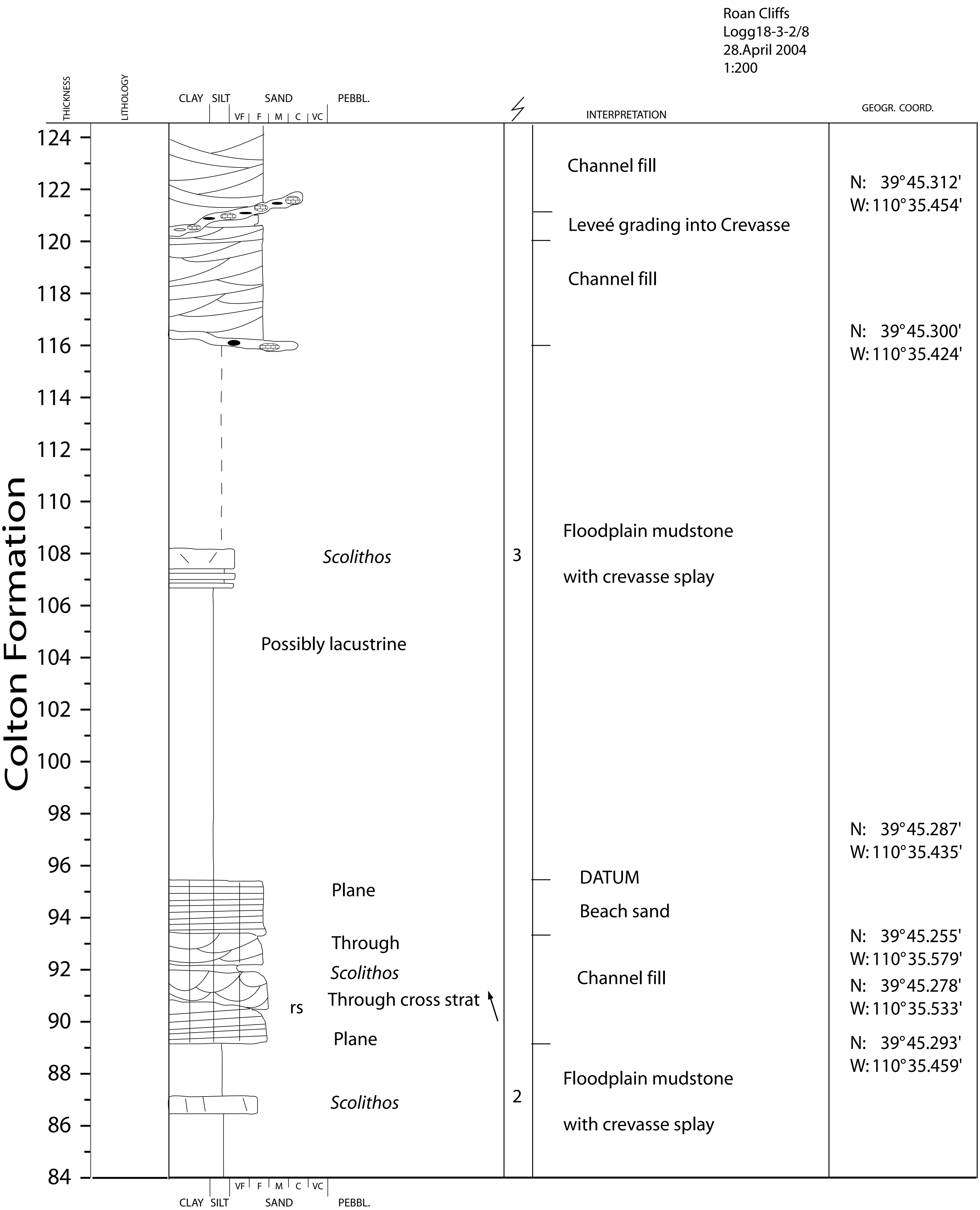
Roan Cliffs
Logg18-2-8/8
24.April 2004
1:200



Logg 18-3

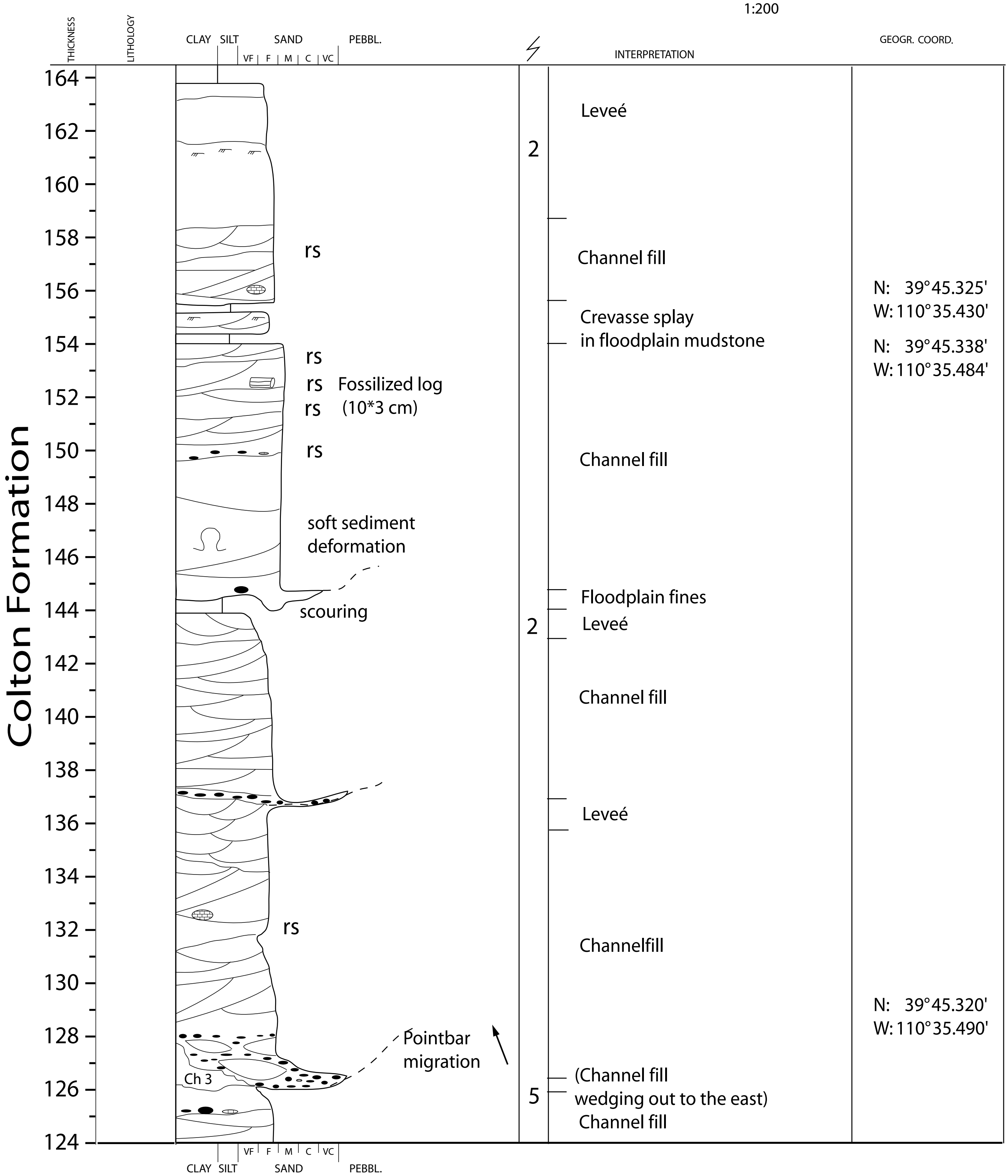


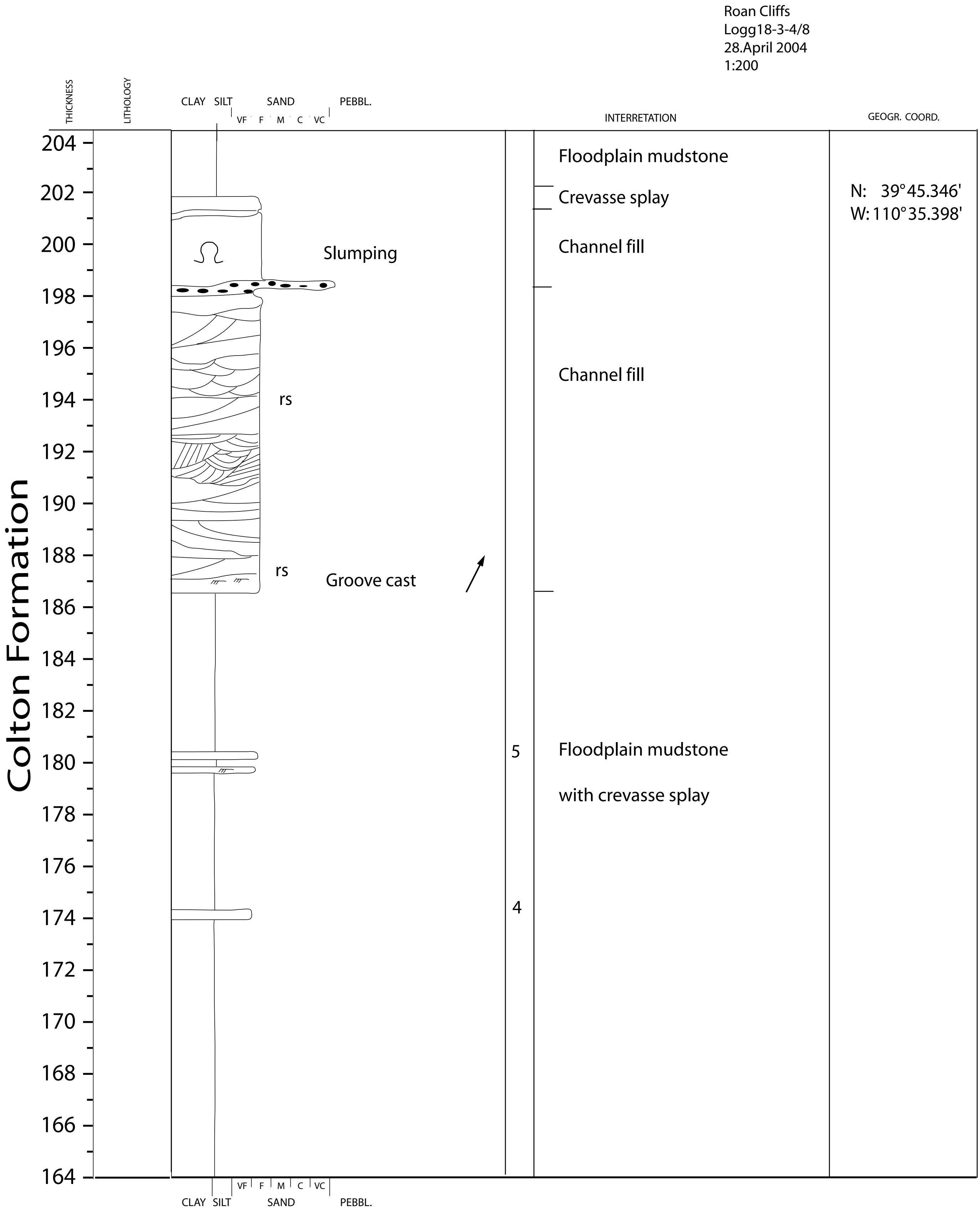
Roan Cliffs
Logg18-3-1/8
28.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATIONF
GEOGR. COORD.
Colton Formation
82
80
78
76
74
72
70
68
66
64
62
60
58
56
10
8
6
4
2
0
4
Floodplain mudstone
with crevasse splay
N: 39°45.251'
W: 110°35.570'
Current ripple
Trough/plane/trough/plane
Parting lination
1
Crevasse splay/Beach sandstone
46 m silt
Floodplain mudstone
with crevasse splay
rs
Channel fill
rs
N: 39°45.206'
W: 110°35.506'
VF F M C VC
CLAY SILT SAND PEBBL.



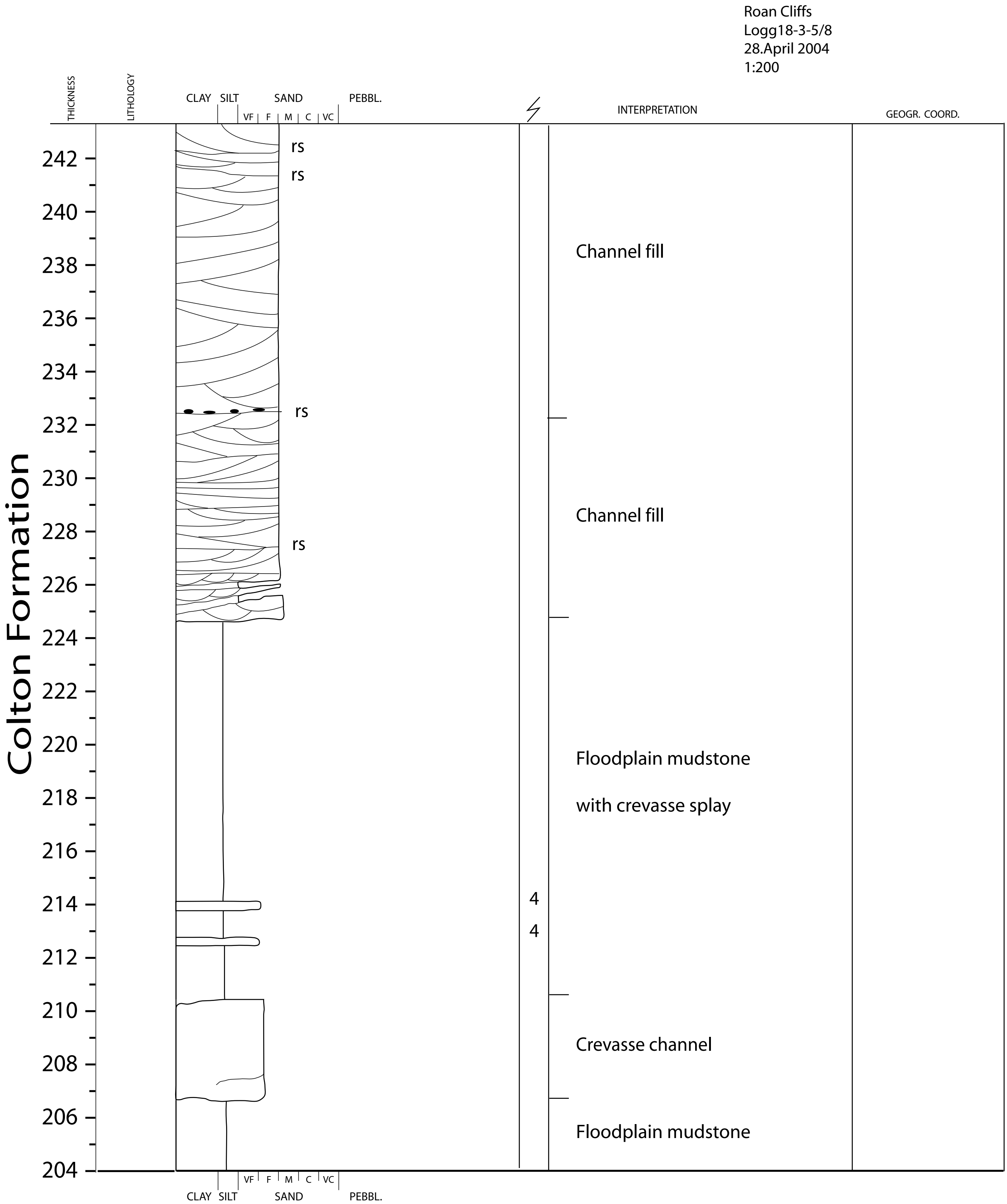
Roan Cliffs
Logg18-3-2/8
28.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATION
GEOGR. COORD.
Colton Formation
124
122
120
118
116
114
112
110
108
106
104
102
100
98
96
94
92
90
88
86
84
Channel fill
Leveé grading into Crevasse
Channel fill
N: 39°45.312'
W: 110°35.454'
N: 39°45.300'
W: 110°35.424'
Scolithos
3
Floodplain mudstone
with crevasse splay
Possibly lacustrine
N: 39°45.287'
W: 110°35.435'
Plane
DATUM
Beach sand
Through
Scolithos
N: 39°45.255'
W: 110°35.579'
Channel fill
rs
Through cross strat
N: 39°45.278'
W: 110°35.533'
Plane
N: 39°45.293'
W: 110°35.459'
Floodplain mudstone
Scolithos
2
with crevasse splay
CLAY SILT SAND PEBBL.
VF F M C VC

Roan Cliffs
Logg18-3-3/8
28.April 2004
1:200

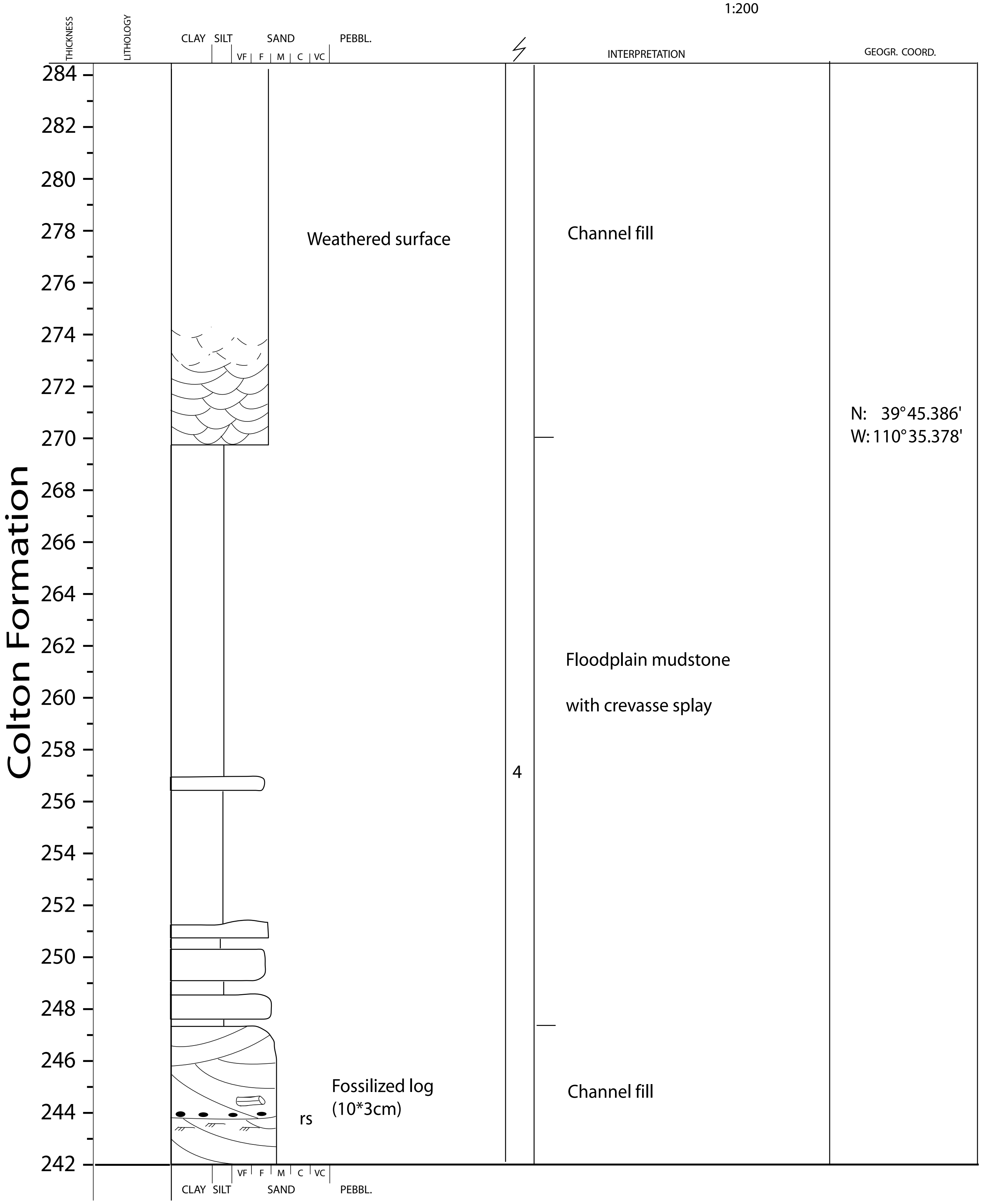




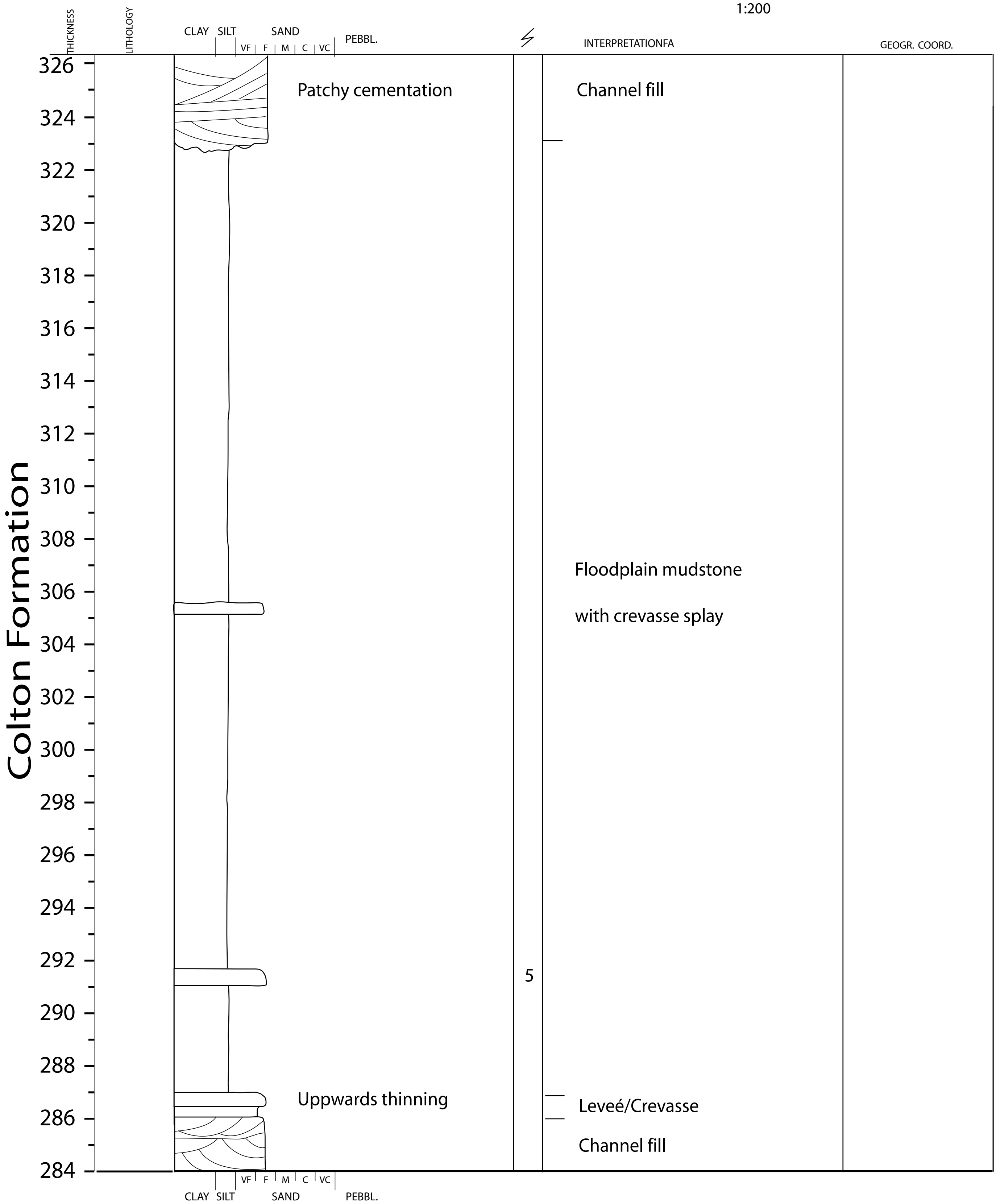
Roan Cliffs
Logg18-3-4/8
28.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERRETATION
GEOGR. COORD.
Colton Formation
204
202
200
198
196
194
192
190
188
186
184
182
180
178
176
174
172
170
168
166
164
Slumping
rs
rs
Groove cast
5
4
Floodplain mudstone
Crevasse splay
Channel fill
Channel fill
Floodplain mudstone
with crevasse splay
N: 39°45.346'
W: 110°35.398'

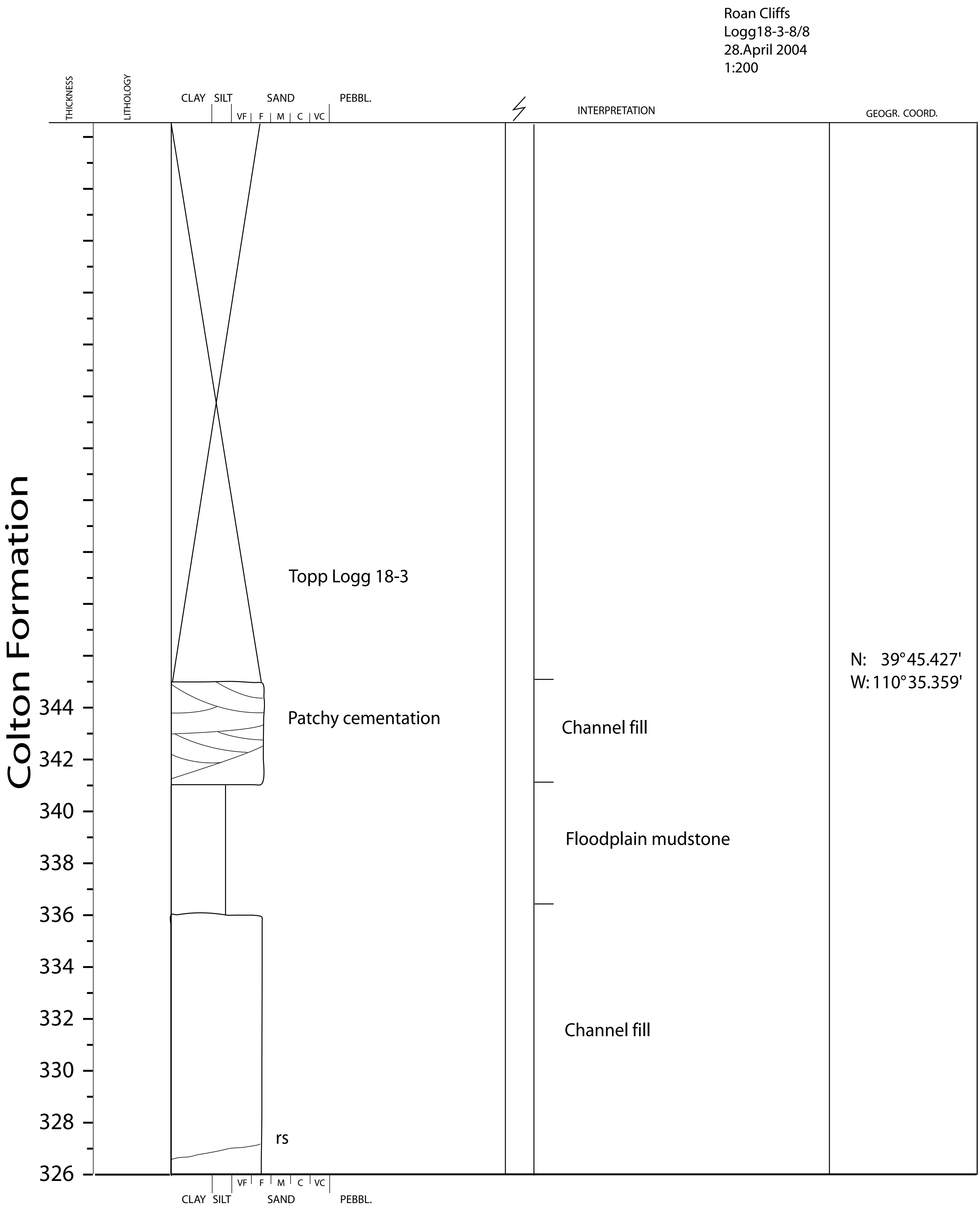

Roan Cliffs
Logg18-3-5/8
28.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATION
GEOGR. COORD.
Colton Formation
242
240
238
236
234
232
230
228
226
224
222
220
218
216
214
212
210
208
206
204
rs
rs
rs
rs
4
4
Channel fill
Channel fill
Floodplain mudstone
with crevasse splay
Crevasse channel
Floodplain mudstone

Roan Cliffs
Logg18-3-6/8
28.April 2004
1:200



Roan Cliffs
Logg18-3-7/8
28.April 2004
1:200




Roan Cliffs
Logg18-3-8/8
28.April 2004
1:200
THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
INTERPRETATION
GEOGR. COORD.
Colton Formation
Topp Logg 18-3
Patchy cementation
rs
Channel fill
Floodplain mudstone
Channel fill
N: 39°45.427'
W: 110°35.359'
344
342
340
338
336
334
332
330
328
326
VF F M C VC
CLAY SILT SAND PEBBL.